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GRAVITY ANOMALIES IN THE UNITED STATES

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ABSTRACT

Gravity data in the United States are used to derive warp anomalies. The correlation of these with tectonic structure is discussed.

In his interesting analysis of gravity anomalies in the Black Hills–Bighorn–Beartooth region Professor R. T. Chamberlin[1] concludes that isostatic anomalies are due to the combined effect of the departures from normal density of superficial rocks and errors in the assumption of perfect isostasy. He has shown that the anomaly is influenced by the elevation of the station because isostatic density is assumed to vary inversely with the elevation. This indicates that the assumption of isostatic compensation is at least partly wrong. In India a correlation of isostatic anomalies with the elevation of the station was detected; and a correction was applied on this account, at first empirically[2] and later by removing from the isostatic anomaly the correction for isostatic compensation which had been applied out to a radius of 120,000 feet.[3] This led to the formu-

[1] "Geologic Analysis of the Gravity Anomalies for the Black Hills–Big Horn–Beartooth Region," *Geol. Soc. Amer. Bull. 46* (1935), pp. 393–408.

[2] E. A. Glennie, "Gravity," *Surv. of India Geodet. Rept.*, Vol. VI (1931), chap. iv, p. 55.

[3] E. A. Glennie, "Gravity Anomalies and the Structure of the Earth's Crust," *Surv. of India Prof. Paper 27* (1932), p. 12.

lation of the crustal-warp hypothesis,[4] according to which isostatic anomalies result from the combined influence of: (1) superficial density anomalies; (2) the effect of deep-seated warpings of the crustal layers; (3) the erroneous assumption of isostasy. It is claimed that warp anomalies employed in India since 1931, with the addition of a "hidden-range" correction, do adequately show the effect of the first two factors only; i.e., the effect of superficial density anomalies and crustal warping within a radius of 120,000 feet of the station. I have recently attempted to show that similar warp anomalies would prove satisfactory in the United States,[5] basing my arguments on approximate warp anomalies obtained by applying an empirical height correction to the isostatic anomalies. Dr. William Bowie has since most kindly placed at my disposal the full data of the isostatic reduction of nearly six hundred gravity stations in the United States, and for these the warp anomalies have been computed. The geologic data given by Professor Chamberlin in his paper can now be used to correct these anomalies, removing the effect of superficial density anomalies. Two corrections are required: (1) geologic correction and (2) topographic density correction. The method employed to obtain them is given below.

GEOLOGIC CORRECTION

This is the correction for density anomalies in the granitic layer below sea-level for which the data to a depth of 8 kilometers have been given. Professor Chamberlin's data extend to a radius of 16 km. for all stations and to a radius of 32 km. in the case of six stations.

In the case of the uncorrected or normal warp anomalies the granitic layer is assumed to have normal density of 2.67. This column, 8 km. deep, exerts an attraction at a station vertically above it as follows: for a column the radius of which is 16 km., the attraction is 0.685 cm./sec^2; for a column with a radius of 32 km., the attraction is 0.786 cm./sec.^{2} Subtracting from this the corresponding geologic attraction given by Chamberlin, the difference is the

[4] E. A. Glennie, "Gravity," *loc. cit.*, p. 48.

[5] "Crustal Warping in the United States," *Gerlands Beiträge zur Geophysik*, Vol. XLVI (1935), pp. 193–97.

geologic correction, shown in Table I. Since the geologic attraction given for all the sixty-five stations is less than normal, the density of the rocks above sea-level is also assumed to be less than normal

TABLE I

	Radius 16 Km.	Radius 32 Km.
	Mountain Stations	
No. 373. Burgess Ranger Station		
F_1 (attraction for normal density)	0.685	0.786
F_2 (geologic attraction)	0.683	0.784
Difference $F_1 - F_2$ (geologic correction)	+ 2 mgals.	+ 2 mgals.
No. 377. East Tensleep, Wyoming:		
F_1	0.685	0.786
F_2	0.683	0.784
$F_1 - F_2$	+ 2 mgals.	+ 2 mgals.
No. 378. Hazelton, Wyoming:		
F_1	0.685	0.786
F_2	0.683	0.784
$F_1 - F_2$	+ 2 mgals.	+ 2 mgals.
	Basin Stations	
No. 366. Gillette, Wyoming:		
F_1	0.685	0.786
F_2	0.648	0.748
$F_1 - F_2$	+37 mgals.	+38 mgals.
No. 367. Ulm, Wyoming:		
F_1	0.685	0.786
F_2	0.642	0.741
$F_1 - F_2$	+43 mgals.	+45 mgals.
No. 431. West of Burlington, Wyoming		
F_1	0.685	0.786
F_2	0.633	0.730
$F_1 - F_2$	+52 mgals.	+56 mgals.

and to the same extent. Though only a makeshift correction, it is believed that it is at least in the right direction and probably not far wrong.

The correction is worked out as follows, taking station No. 368 as an example, since this has the lowest density for the 8-km. column:

At station No. 368, Buffalo, Wyoming, the geologic attraction is 0.630 and the normal attraction, 0.685. Therefore

$$\text{Density} = 2.67 \times \left(\frac{0.630}{0.685}\right)$$
$$= 2.67(1 - 0.0801),$$

whence 0.0801 × normal topographic correction is the correction required for this station. By referring to the reduction data for this station, the result is as given in Table II. Thus the topographic density correction to the warp anomaly is +13 mgals. The result above has been given out to an extreme radius so as to show how rapidly the effect falls off. The greater part of the topographic correction is

TABLE II

STATION 368

	Outer Radius				
	12.4 Km.	18.8 Km.	28.8 Km.	58.8 Km.	166.7 Km.
Total normal topographic correction	+0.1514	+0.1550	+0.1571	+0.1588	+0.1603
Topographic density correction	+0.0121	+0.0124	+0.0126	+0.0127	+0.0128

usually within the first 12 km.; and the geology outside this can, as a rule, be neglected without introducing any serious error.

The resulting anomalies are given in Table III as follows: "normal warp anomaly" contains no geologic or topographic density corrections; "first corrected warp anomaly" is the normal warp anomaly plus geologic correction; and "final warp anomaly" is the first corrected anomaly plus topographic density correction. The table shows also the normal warp anomalies for some additional gravity stations in the same area, but the final warp anomalies for these cannot be given in the absence of geologic data.

The equation for theoretical gravity used in the computation of the isostatic anomalies is the international formula,

$$\gamma_0 = 978.049(1 + .005287 \sin^2 \phi - .000006 \sin^2 2\phi).$$

The depth of compensation assumed is 113.7 km.

Here the terms within the bracket are based on theoretical considerations; but the factor 978.049, which is the assumed equatorial value of gravity, has been obtained empirically from a consideration

TABLE III

No.	Station	Normal Warp Anomaly	First Corrected Warp Anomaly	Final Warp Anomaly	No.	Station	Normal Warp Anomaly	First Corrected Warp Anomaly	Final Warp Anomaly
357	Box Elder, S.D.	+ 5	+19	+21	391	Midwest, Wyo.	−19	− 6	− 3
358	Rapid City, S.D.	+26	+30	+31	392	Kaycee, Wyo.	−19	− 6	− 3
359	Pactola, S.D.	+28	+30	+31	393	Sussex, Wyo.	−34	+11	+21
360	Tigerville, S.D.	+24	+26	+27					
361	Smith Ranch, S.D.	+16	+18	+19	421	Big Horn, Wyo.	−18	+17	+24
362	Provo, S.D.	+31	+40	+42	422	Southeast of Sheridan, Wyo.	−25	+20	+29
363	Moon, S.D.	+24	+27	+28					
364	Osage, Wyo.	+ 4	+11	+12	423	Experiment Farm, Wyo.	−24	+16	+24
365	Kara, Wyo.	+ 3	+23	+27					
366	Gillette, Wyo.	0	+37	+45	424	Veterans' Hospital, Wyo.	−16	+22	+29
367	Ulm, Wyo.	−42	+ 1	+10	425	Beckton, Wyo.	−31	+ 4	+11
368	Buffalo, Wyo.	−38	+17	+30	426	Durkee, Wyo.	−21	+23	+31
369	Acme Mine, Wyo.	−14	+19	+25	427	Greybull, Wyo.	−28	+12	+20
370	Ranchester, Wyo.	−19	+11	+17	428	West of Otto, Wyo.	−43	+ 5	+15
371	Dayton, Wyo.	−25	−13	−11	429	Cody, Wyo.	−60	−47	−44
372	Steamboat Point, Wyo.	+15	+17	+18	430	Southeast of McCullock, Wyo.	−60	− 9	+ 3
373	Burgess Ranger Station, Wyo.	+17	+19	+20	431	West of Burlington, Wyo.	−54	− 2	+10
374	Medicine Mountain, Wyo.	+25	+27	+28	432	Pezan Ranch, Wyo.	−64	−14	− 2
375	Five Springs Bench, Wyo.	0	+ 9	+11	433	East of Meeteetse, Wyo.	−66	−24	−13
376	Ten Sleep, Wyo.	+27	+34	+35	434	Meeteetse, Wyo.	−42	−17	−10
					435	Irma Flats, Wyo.	−36	−24	−21
377	East Ten Sleep, Wyo.	+25	+27	+28	436	Oregon Basin, Wyo.	−56	−43	−40
					437	Wymont, Wyo.	− 2	0	+ 1
378	Hazelton, Wyo.	+21	+23	+24	438	Red Lodge, Mont.	−34	+ 2	+12
379	South Fork Inn, Wyo.	+32	+34	+35	439	Bridger, Mont.	+20	+34	+36
					440	Bowler, Mont.	−15	− 9	− 8
380	Kane, Wyo.	−13	− 2	0					
381	Lovell, Wyo.	−34	−20	−17	441	Gardiner, Mont.	−47	−20	−12
382	Garland, Wyo.	−50	−22	−17	443	Clark Fork River, Wyo.	− 7	− 5	− 4
383	Eagle Siding, Wyo.	−60	−12	− 1					
384	Elk Basin, Wyo.	−47	−34	−31	444	Arvada, Wyo.	−38	− 2	+ 4
385	South of Basin, Wyo.	−24	− 3	+ 1	445	Spotted Horse, Wyo.	− 4	+33	+40
386	Rairden, Wyo.	− 9	+24	+30	446	Wildcat Creek, Wyo.	− 4	+31	+38
387	Thermopolis, Wyo.	−15	− 8	− 7	447	Rozet, Wyo.	−14	+17	+24
388	Worland, Wyo.	−20	+33	+44	448	Northeast of Gillette, Wyo.	+ 5	+39	+47
389	Winchester, Wyo.	−35	−11	− 6					
390	Nine Mile Lake, Wyo.	−30	− 9	− 4	449	Southeast of Gillette, Wyo.	−14	+24	+32

of gravity results corrected for topography and isostatic compensation.

In the case of the normal warp anomaly the correction for isostatic compensation in the inner zones has been removed from the reduced gravity results; hence a new empirical value for equatorial gravity is required. To derive this, preliminary warp anomalies were obtained using the international formula for all gravity sta-

tions in the United States, and from these the average anomaly for each degree square in which there are stations was secured. As these squares are rather unevenly distributed, the values so obtained were again averaged by 2° squares, and the mean value of all these 2° squares taken for the final result.

The mean anomaly derived in this way amounted to −21 mgals. Assuming that the squares are evenly distributed over up- and downwarps of the crust, a correction of +21 mgals. is required to all warp anomalies, i.e., the equatorial value of gravity employed should be 978.028 instead of 978.049. Hence the theoretical gravity formula used for all three classes of warp anomalies in Table III and for normal warp anomalies in Table IV is

$$\gamma_0 = 978.028(1 + .0005287 \sin^2 \phi - .000006 \sin^2 2\phi) .$$

If geologic data were available and applied to the results for all stations, the same procedure should be repeated and a correction applied to the final warp anomalies in Table III.

Figure 1 (p. 777) shows the distribution of gravity anomalies in this area: (*A*) isostatic anomalies; (*B*) normal warp anomalies; (*C*) final warp anomalies with the addition of a constant correction of −14 mgals.

I think it will be agreed that (*B*) is an improvement on (*A*) and that (*C*) is better than (*B*). In fact, the importance of including the geologic data pointed out by Chamberlin is clear. Some idea of numerical values can be obtained from the sections on Lat. 44°00′ and Lat. 44°30′ in Figure 2 (p. 778).

The Black Hills and Powder River areas appear to be broadly upwarped areas of the crust, to the west of which crustal warpings become increasingly abrupt. An approximate value for the amount of crustal warping can be obtained from Table III in *Professional Paper 27* of the Survey of India. Thus an anomaly of +22 mgals. in the Bighorn Mountains indicates an underlying crustal upwarp of about 6,000 feet; and the maximum negative anomaly, −58 mgals., in the Bighorn Basin corresponds to a downwarp of about 18,000 feet, the average downwarp in this basin being about 10,000 feet (−35 mgals.). For other parts of the United States similar geologic data are not available; so the final warp anomalies cannot be prepared.

TABLE IV

Station No.	Latitude	Longitude	Normal Warp Anomaly	Station No.	Latitude	Longitude	Normal Warp Anomaly
1	24°33′.6	81°48′.4	+22	51	44°44′.2	110°42′.0	−30
2	26 42.8	80 02.8	+36	52	44 33.4	110 48.1	−49
3	26 56.2	82 03.0	+26	53	47 39.6	122 18.3	−73
4	29 43.5	84 58.8	+18	54	37 47.5	122 25.7	− 6
5	29 57.0	90 04.2	+ 4	55	37 20.4	121 38.6	0
6	32 28.0	91 45.0	+34	56	47 36.5	122 19.8	−74
7	29 18.2	94 47.5	+ 8	57	46 05.4	88 38.4	+44
8	26 04.7	97 12.4	+44	58	47 48.6	92 01.0	+31
9	27 30.5	99 31.2	− 6	59	48 58.1	97 14.9	+33
10	30 16.5	97 44.3	+ 4	60	43 41.8	98 01.8	+ 9
11	30 17.2	97 44.2	+ 2	61	32 28.4	100 24.1	−30
12	34 56.2	95 46.2	−14	62	30 01.3	99 07.6	+33
13	34 45.0	92 16.4	+48	63	31 46.3	106 29.0	−11
14	35 36.7	87 02.5	+40	64	31 21.3	110 56.6	−65
15	33 45.0	84 23.3	−12	65	32 43.3	114 37.0	+24
16	33 54.8	82 18.0	+32	66	33 53.4	118 13.2	−33
17	32 47.2	79 56.0	− 3	67	37 42.2	117 14.5	−45
18	34 43.1	76 39.8	− 3	68	36 03.9	112 07.1	−34
19	38 02.0	78 30.3	+ 1	69	36 05.3	112 06.8	−43
20	39 25.0	79 19.8	+ 7	70	35 31.8	108 44.2	−58
21	38 53.2	77 00.5	+56	71	35 35.8	105 12.1	−42
22	38 53.3	77 01.5	+58	72	35 12.8	100 11.4	+29
23	39 17.8	76 37.3	+ 7	73	33 45.3	96 32.8	+16
24	39 57.1	75 11.7	+42	74	44 58.7	93 13.9	+73
25	40 21.0	74 39.5	− 1	75	44 21.1	103 45.6	+27
26	40 44.0	74 02.0	+44	76	46 48.5	100 47.0	+ 6
27	40 48.5	73 57.7	+42	77	48 23.8	107 05.3	+31
28	42 16.5	71 48.5	− 4	78	48 16.4	116 33.3	− 4
29	42 21.6	71 03.8	+25	79	43 37.2	116 12.3	− 2
30	42 22.8	71 07.8	+24	80	46 11.3	123 50.2	+ 7
31	45 11.2	67 16.9	+13	81	41 18.3	122 19.6	−30
32	42 27.1	76 29.0	−11	82	41 35.1	109 13.2	−30
33	41 30.4	81 36.6	+10	83	41 07.4	101 21.3	−13
34	39 08.3	84 25.3	− 6	84	38 56.3	77 04.0	+53
35	39 28.7	87 23.8	+ 6	85	44 49.1	73 17.5	+20
36	41 47.4	87 36.1	+10	86	44 17.5	73 59.1	+ 8
37	43 04.6	89 24.0	+ 8	87	44 40.1	74 58.8	+36
38	38 38.0	90 12.2	+10	88	43 18.4	78 49.6	+ 8
39	39 05.8	94 35.4	− 4	89	45 03.8	83 27.0	− 6
40	38 43.7	98 13.5	+19	90	36 50.5	75 58.4	−29
41	38 54.7	101 35.4	−25	91	36 00.2	78 53.5	+53
42	38 50.7	104 49.0	−51	92	30 40.2	81 27.7	+28
43	38 50.3	105 02.0	−41	93	30 49.2	88 20.5	−26
44	39 40.6	104 56.9	−45	94	33 07.6	88 10.8	0
45	38 32.6	106 56.0	−39	95	36 35.5	89 31.6	+20
46	39 04.2	108 33.9	− 8	96	34 35.2	94 14.6	−43
47	38 59.4	110 09.9	−43	97	31 36.2	94 37.8	+ 5
48	39 50.8	111 00.8	−48	98	30 21.5	103 39.7	− 3
49	40 46.1	111 53.8	−19	99	34 23.2	103 01.8	−34
50	44 43.3	110 29.7	−54	100	36 40.7	101 28.7	−25

TABLE IV—*Continued*

Station No.	Latitude	Longitude	Normal Warp Anomaly	Station No.	Latitude	Longitude	Normal Warp Anomaly
101	36°25′.9	84°32′.6	+45	151	35°13′.8	80°50′.8	+36
102	36 06.2	82 07.9	− 7	152	35 35.9	82 33.3	− 9
103	36 08.5	82 07.2	−39	153	35 09.4	84 52.9	−15
104	38 20.9	81 37.7	−14	154	36 06.1	80 17.0	−29
105	40 47.9	77 51.8	−13	155	35 57.7	83 55.0	−13
106	47 14.9	68 36.0	+ 3	156	36 35.4	82 12.0	−15
107	45 32.6	90 17.8	+31	157	25 28.4	80 28.9	−19
108	46 17.2	96 05.0	+ 5	158	27 30.2	81 27.0	0
109	44 48.0	106 58.7	+ 3	159	28 36.7	80 48.4	+16
110	46 14.2	112 07.3	−49	160	28 48.6	81 53.0	+ 3
111	47 42.4	121 22.3	−33	161	29 08.3	83 02.1	− 3
112	47 03.4	122 52.7	+54	162	32 49.8	83 38.0	+35
113	45 21.4	119 33.2	−24	163	31 34.3	84 09.0	+18
114	39 19.6	120 11.4	−70	164	30 24.5	87 12.9	+ 3
115	40 58.4	117 43.8	−29	165	32 38.5	85 22.8	−16
116	39 14.9	114 53.4	−67	166	34 43.8	86 35.2	−13
117	42 16.1	104 44.0	+17	167	33 36.5	91 12.2	+ 5
118	44 21.9	100 20.8	+20	168	35 08.8	90 03.3	+29
119	42 30.8	94 11.4	+26	169	36 29.3	91 27.0	+25
120	41 06.4	90 57.0	+ 5	170	36 51.6	87 28.0	+19
121	42 58.0	85 40.8	+15	171	37 38.9	84 46.4	−19
122	41 37.7	85 00.6	+24	172	37 49.1	79 49.6	−31
123	42 39.1	73 46.1	−27	173	31 49.4	86 38.0	+ 4
124	41 22.4	74 41.1	−22	174	33 30.8	86 48.8	−21
125	39 21.9	74 25.0	− 3	175	37 47.2	79 26.6	−17
126	40 56.0	72 18.4	− 3	176	37 40.6	82 45.6	−14
127	41 40.7	69 57.3	+ 6	177	44 45.8	85 37.2	+17
128	44 06.3	69 06.9	+ 5	178	46 20.8	85 57.6	+16
129	44 29.5	71 34.3	−10	179	44 53.2	87 52.0	− 9
130	43 33.0	73 23.8	−25	180	44 23.6	89 46.0	−31
131	43 02.7	74 51.2	−13	181	44 03.2	91 38.4	+27
132	43 58.3	75 54.6	−11	182	44 57.8	92 23.0	−38
133	42 03.7	76 48.6	−22	183	45 32.4	92 00.0	−38
134	42 07.8	80 04.8	−14	184	45 34.0	93 11.0	−16
135	39 16.0	81 33.7	−13	185	46 21.3	94 12.1	+22
136	39 57.8	82 59.4	0	186	45 27.5	98 29.0	+20
137	39 45.9	86 08.8	+13	187	45 01.3	102 04.0	+14
138	39 47.7	89 39.5	− 1	188	46 18.4	103 53.0	+30
139	37 41.1	92 39.1	+20	189	48 20.3	100 26.0	+40
140	37 05.4	94 30.8	+25	190	48 54.7	103 19.0	+21
141	35 23.3	94 25.5	− 3	191	47 46.2	96 36.0	+24
142	33 25.5	94 02.5	+27	192	48 06.8	105 12.0	+21
143	34 30.1	93 03.6	+32	193	46 24.2	105 50.0	+26
144	31 18.6	92 26.0	+12	194	45 54.0	108 19.6	+ 1
145	31 41.5	89 02.3	+30	195	42 50.0	108 43.0	−19
146	37 32.2	77 26.1	+21	196	44 17.8	93 15.0	+48
147	36 40.2	77 31.0	+33	197	43 58.6	94 36.0	+17
148	35 36.8	77 22.3	+ 1	198	43 17.7	103 49.2	+40
149	34 14.2	77 56.6	−12	199	44 55.8	96 01.0	+28
150	34 42.0	79 54.0	+20	200	45 04.5	94 12.0	+19

TABLE IV—*Continued*

Station No.	Latitude	Longitude	Normal Warp Anomaly	Station No.	Latitude	Longitude	Normal Warp Anomaly
201	44°04′.2	102°25′.0	+30	251	34°26′.9	120°28′.3	−15
202	44 15.5	104 58.0	+ 4	252	35 10.6	120 43.9	−15
203	46 47.0	92 06.4	+62	253	36 07.8	121 01.2	−21
204	43 16.8	92 47.0	−16	254	36 36.0	121 53.8	− 8
205	42 23.0	97 19.0	+ 8	255	36 51.1	121 24.0	−39
206	42 52.3	100 31.0	+14	256	37 26.6	122 09.7	− 4
207	40 04.0	80 43.4	−18	257	37 19.4	122 23.3	−12
208	40 44.6	93 43.0	+ 2	258	38 04.0	122 48.7	− 1
209	39 06	76 51.0	+51	259	38 27.3	123 03.8	−16
210	40 10.(	76 53.1	−14	260	39 16.8	80 20.3	+ 2
211	40 27.4	80 00.6	−13	261	39 20.8	79 40.2	+ 5
212	39 04.9	77 08.8	+62	262	39 26.9	79 32.9	+ 9
213	38 49.0	76 45.2	+31	263	39 25.6	79 29.4	+ 5
214	38 47.7	77 19.6	+53	264	39 23.6	79 10.4	− 3
215	37 58.8	75 50.7	−10	265	40 20.0	74 47.6	−19
216	38 18.1	77 27.5	+23	266	40 48.1	74 12.2	− 3
217	39 09.7	75 32.0	+10	267	40 20.0	74 35.8	+ 1
218	47 15.8	88 27.6	+45	268	41 44.8	72 41.8	+26
219	39 38.5	77 43.5	−36	269	43 55.9	103 34.1	+15
220	48 45.1	122 28.6	−24	270	43 51.4	104 11.8	+ 7
221	47 58.6	122 12.5	−54	271	41 40.1	103 05.9	−23
222	47 31.6	122 02.2	−30	272	41 07.4	105 18.3	− 3
223	48 07.0	123 25.9	+ 6	273	40 01.2	105 16.6	−53
224	48 06.8	122 45.6	− 2	274	39 59.8	105 05.2	−51
225	47 51.3	122 35.1	+22	275	39 59.3	104 48.9	−33
226	47 34.4	122 37.8	− 5	276	39 44.6	105 31.1	−40
227	47 15.2	122 26.3	+ 8	277	30 13.8	88 01.2	−13
228	47 14.6	124 13.4	−45	278	30 02.8	88 52.3	−14
229	45 31.4	122 40.7	+11	279	29 01.0	89 09.8	0
230	45 27.4	123 50.6	+44	280	28 58.0	89 22.8	+37
231	44 38.2	124 03.3	+28	281	29 21.8	89 27.9	+31
232	44 02.7	123 05.6	+21	282	29 34.7	89 47.7	+29
233	43 22.0	124 13.4	+23	283	29 02.8	90 21.3	− 8
234	42 44.2	123 25.8	+13	284	29 41.7	91 12.5	+19
235	40 01.6	122 07.2	+13	285	29 43.4	93 52.0	− 5
236	38 34.8	121 29.8	+ 6	286	30 05.2	94 06.0	+ 3
237	39 24.9	123 22.0	−38				
238	40 48.2	124 09.7	−30	296	39 04.4	97 04.5	−15
239	38 26.4	122 43.0	−14	297	39 09.7	96 25.6	−28
240	32 42.8	117 09.9	−14	298	39 12.5	96 05.2	−12
				299	39 50.4	96 03.6	−23
				300	34 22.1	96 25.5	+32
241	33 11.6	117 22.5	+ 3	301	34 19.6	96 46.6	+73
242	34 07.5	117 12.5	−60	302	34 13.3	97 07.0	+ 4
243	34 03.2	117 45.2	−27	303	34 14.1	97 30.8	+24
244	33 46.3	118 11.6	− 3	304	34 11.4	97 32.6	+30
245	33 50.6	118 23.3	−10	305	34 18.3	97 31.6	+ 3
246	34 11.1	118 18.9	−15	306	32 38.6	95 39.6	+19
247	34 35.0	118 08.0	−37	307	32 38.6	95 37.0	+14
248	35 03.2	118 10.4	−49	308	32 38.6	95 42.2	+20
249	35 03.8	119 24.0	−59	309	30 34.1	97 24.5	+10
250	34 16.8	119 17.6	−66	310	30 38.0	97 40.1	− 1

TABLE IV—*Continued*

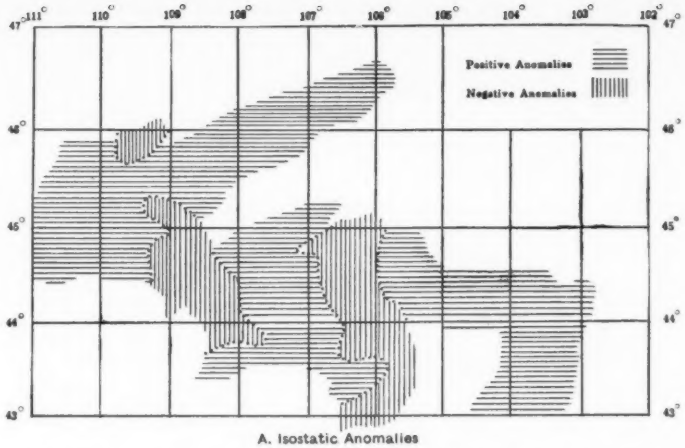
Station No.	Latitude	Longitude	Normal Warp Anomaly	Station No.	Latitude	Longitude	Normal Warp Anomaly
311	29°19′.5	95°42′.7	− 2	361	43°47′.3	103°30′.7	+16
312	29 17.0	95 43.7	− 4	362	43 11.6	103 49.4	+31
313	29 15.5	95 45.2	− 6	363	43 56.0	103 59.9	+24
314	34 08.1	118 07.6	− 9	364	43 58.9	104 24.9	+ 4
315	34 13.4	118 03.4	−14	365	44 11.8	104 51.0	+ 3
316	41 17.5	82 13.2	+ 3	366	44 17.5	105 29.5	0
317	35 47.5	78 40.3	+12	367	44 39.2	106 35.5	−42
318	35 30.4	117 39.3	−18	368	44 18.8	106 41.3	−38
319	42 57.1	78 49.3	− 3	369	44 54.4	107 00.5	−14
320	25 42.8	80 15.1	+12	370	44 54.5	107 09.5	−19
321	38 49.4	77 01.5	+55	371	44 51.7	107 17.3	−25
322	39 08.2	77 11.9	+54	372	44 47.4	107 23.7	+15
323	38 31.5	77 17.2	+32	373	44 46.4	107 30.9	+17
324	37 15.6	76 40.8	− 1	374	44 48.8	107 52.0	+25
325	37 16.3	76 42.6	+ 5	375	44 48.3	108 08.0	0
326	37 49.5	78 00.2	+27	376	44 02.4	107 26.0	+27
327	38 01.8	78 51.5	− 7	377	44 10.1	107 13.0	+25
328	38 13.9	79 11.0	−28	378	44 08.2	106 55.6	+21
329	38 23.5	79 33.0	−24	379	44 17.4	106 56.7	+32
330	38 36.0	79 50.9	−17	380	44 51.0	108 11.6	−13
331	38 53.1	79 50.7	− 8	381	44 50.5	108 25.4	−34
332	39 16.4	81 03.2	− 3	382	44 46.6	108 39.1	−50
333	39 51.3	79 39.6	− 2	383	44 40.4	108 55.2	−60
334	39 47.6	79 10.0	+ 5	384	44 59.5	108 52.8	−47
335	39 37.0	78 15.6	−19	385	44 20.1	108 01.7	−24
336	39 25.7	77 30.0	+27	386	44 11.2	107 53.0	− 9
337	38 53.6	77 02.0	.....	387	43 35.7	108 11.8	−15
338	38 56.6	77 03.4	+54	388	44 00.8	107 59.3	−20
339	38 03.8	78 39.0	− 9	389	43 53.0	108 08.9	−35
340	32 56.4	85 57.0	−25	390	42 59.2	106 19.9	−30
341	33 02.8	86 05.5	−11	391	43 24.7	106 16.0	−19
342	33 09.3	86 15.0	−10	392	43 42.8	106 37.9	−19
343	33 32.6	86 33.5	−15	393	43 41.9	106 16.9	−34
344	33 25.7	86 27.1	− 9	394	41 06.9	83 10.0	+ 1
345	33 17.1	86 20.5	− 1	395	40 31.9	81 28.4	+ 8
346	33 36.3	86 57.6	−10	396	39 09.9	78 09.6	−32
347	33 43.8	87 04.6	+13	397	38 40.1	78 28.8	−24
348	32 27.5	88 20.0	−15	398	29 40.4	82 20.1	− 1
349	32 21.0	88 18.5	− 3	399	30 46.5	85 14.1	−15
350	32 15.5	88 05.5	+ 8	400	30 26.8	91 09.6	−11
351	32 12.5	87 58.5	− 3	401	29 41.4	95 25.1	− 8
352	32 12.0	87 46.0	+ 3	402	29 40.1	95 23.5	− 3
353	32 08.0	87 40.0	− 3	403	29 35.8	95 29.5	− 3
354	32 03.0	87 34.0	− 1	404	29 31.4	95 34.6	− 4
355	31 58.0	87 30.5	+ 3	405	29 19.1	95 42.3	− 1
356	31 56.5	87 21.0	+ 6	406	30 05.4	94 07.6	− 7
357	44 06.8	103 03.3	+ 5	407	30 20.6	95 05.5	− 6
358	44 05.7	103 16.9	+26	408	33 24.1	101 02.8	+42
359	44 04.1	103 29.1	+28	409	33 23.7	101 10.5	+28
360	43 59.6	103 41.1	+24	410	33 24.6	101 16.4	+15

TABLE IV—*Continued*

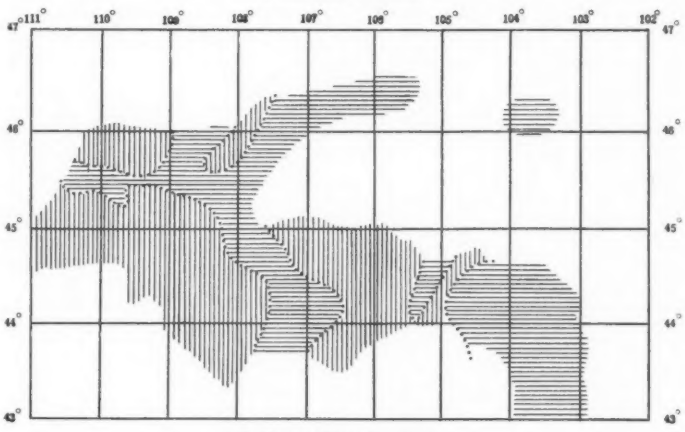
Station No.	Latitude	Longitude	Normal Warp Anomaly	Station No.	Latitude	Longitude	Normal Warp Anomaly
411	33°24′.0	101°22′.2	− 7	461	24°37′.3	81°31′.8	+27
412	33 37.0	100 59.8	0	462	24 51.8	80 43.0	+14
413	33 44.6	101 02.3	−27	463	25 01.2	80 30.3	+10
414	33 25.5	100 55.3	+14	464	25 08.1	80 24.1	+ 7
415	33 24.6	100 43.0	+ 7	465	25 16.5	80 18.4	+ 5
416	33 22.3	100 38.6	+15	466	25 17.8	80 39.8	− 2
417	33 11.5	101 01.3	+ 7	467	25 13.2	80 52.1	+16
418	32 59.3	101 07.1	− 6	468	25 46.2	80 31.3	+19
419	33 22.9	101 33.9	−22	469	25 46.0	80 18.0	+16
420	38 52.2	99 20.5	− 9	470	25 45.9	80 42.0	+12
421	44 40.8	106 59.5	−18	471	25 47.6	80 52.4	+ 9
422	44 39.6	106 50.5	−25	472	25 45.6	81 02.3	+ 6
423	44 50.4	106 50.3	−24	473	25 52.0	81 05.6	− 7
424	44 49.7	106 59.3	−16	474	25 54.1	81 17.8	− 6
425	44 44.9	107 07.9	−31	475	26 09.6	81 20.4	+ 2
426	44 06.9	107 54.9	−21	476	26 03.8	81 42.0	+13
427	44 25.0	108 11.0	−28	477	25 55.9	81 42.5	+ 9
428	44 24.9	108 21.1	−43	478	26 20.4	81 46.2	+ 3
429	44 33.0	109 01.7	−60	479	26 35.6	81 44.1	−15
430	44 27.6	108 37.2	−60	480	27 06.2	81 56.0	+ 8
431	44 25.9	108 31.3	−54	481	27 15.8	81 53.1	+17
432	44 22.1	108 39.6	−64	482	27 07.1	81 20.5	+25
433	44 17.8	108 45.4	−66	483	27 12.8	81 26.7	+30
434	44 08.9	108 52.4	−42	484	27 12.7	81 39.7	+15
435	44 27.1	109 09.1	−36	485	27 12.0	81 20.0	+29
436	44 25.2	108 54.7	−56	486	27 12.8	81 11.7	+15
437	44 58.9	109 24.9	− 2	487	27 14.1	80 59.9	−15
438	45 11.1	109 12.6	−34	488	26 54.5	80 36.7	+14
439	45 20.2	108 55.6	+20	489	27 18.1	80 40.8	− 9
440	45 12.0	108 46.0	−15	490	27 26.4	80 20.9	+13
441	45 01.6	110 42.2	−47	491	27 22.2	80 30.2	+ 4
442	44 55.0	110 15.9	−41	492	27 50.6	81 31.7	+ 3
443	44 56.7	109 47.0	− 7	493	28 04.3	81 47.2	−13
444	44 39.1	106 07.5	−38	494	28 34.4	81 58.7	+ 7
445	44 42.6	105 50.3	− 4	495	29 11.9	82 08.1	+ 1
446	44 29.4	105 35.3	− 4	496	28 21.7	82 13.6	− 1
447	44 16.3	105 11.1	−14	497	27 59.6	82 13.9	+22
448	44 20.1	105 24.8	+ 5	498	26 27.2	82 00.9	+16
449	44 08.8	105 24.3	−14	499	26 08.5	81 47.5	+19
450	42 44.8	108 02.0	−18	500	26 00.0	81 36.0	+10
451	41 59.9	103 40.0	−34	501	26 00.4	81 20.4	− 2
452	41 19.7	102 23.1	− 9	502	25 51.0	80 56.8	0
453	41 04.8	100 57.5	− 2	503	25 50.8	81 23.3	+ 5
454	42 57.6	100 31.7	+ 2	504	26 29.0	81 32.4	− 4
455	40 02.7	102 03.2	−16	505	26 25.4	81 11.3	0
456	37 52.5	100 52.9	−25	506	26 25.6	81 02.9	+ 7
457	41 09.3	81 21.3	+ 9	507	26 37.2	81 26.6	+10
458	27 14.1	80 50.1	−21	508	26 25.6	81 25.2	+ 6
459	25 36.3	80 20.8	− 6	509	25 38.4	80 29.5	+11
460	24 40.1	81 20.4	+25	510	25 54.6	80 23.5	+19

TABLE IV—*Continued*

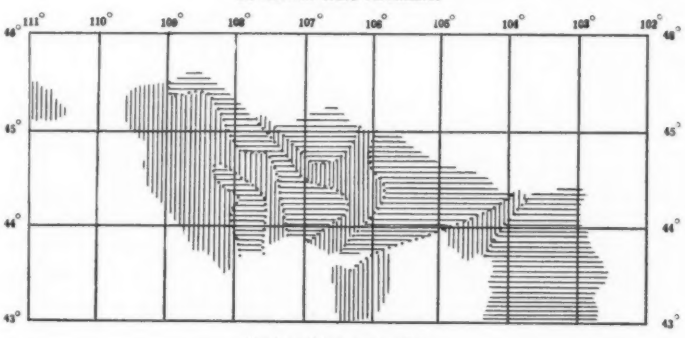
Station No.	Latitude	Longitude	Normal Warp Anomaly	Station No.	Latitude	Longitude	Normal Warp Anomaly
511	25°59′.1	80°07′.1	+22	549	42°04′.9	72°28′.9	+11
512	41 19.1	72 55.4	+36	550	42 19.0	72 38.2	+21
513	41 19.5	72 56.9	+39	551	42 21.0	72 41.6	+12
514	41 21.6	72 59.9	+49	552	42 22.3	72 31.0	+18
515	41 18.2	72 59.5	+52	553	42 45.8	71 27.5	+18
516	41 21.1	72 58.5	+48	554	44 56.7	109 35.8	−11
517	41 16.6	72 50.5	+28	555	45 16.8	109 14.3	+ 3
518	41 16.6	72 47.4	+27	556	45 21.2	109 10.7	+12
519	41 19.2	72 45.3	+30	557	45 17.4	109 25.7	− 7
520	41 19.7	72 48.8	+30	558	45 21.2	109 30.3	+ 6
521	41 27.5	72 41.8	+27	559	44 27.9	109 31.4	+ 8
522	41 25.7	72 38.9	+21	560	45 13.4	110 53.4	−25
523	41 18.8	72 52.6	+30	561	45 32.4	110 33.9	+ 6
524	41 34.6	72 38.1	+23	562	45 25.4	110 33.9	−25
525	41 34.0	72 33.6	+15	563	45 39.5	110 33.1	− 9
526	41 38.7	72 39.1	+24	564	45 42.9	110 23.8	−16
527	41 49.7	72 44.0	+33	565	45 45.0	110 12.5	−13
528	41 34.0	72 53.9	+39	566	45 35.7	110 11.2	+ 2
529	41 33.4	72 58.0	+49	567	45 42.7	109 33.2	−24
530	41 40.5	72 54.1	+40	568	45 39.9	108 46.3	+ 7
531	41 41.0	72 58.1	+41	569	45 38.4	109 15.0	− 2
532	41 50.4	72 48.9	+30	570	45 30.9	109 26.5	+ 5
533	41 49.7	72 54.1	+42	571	45 41.5	108 21.1	− 6
534	41 49.6	72 39.3	+23	572	45 43.2	107 36.4	+11
535	41 47.2	72 26.7	+ 5	573	45 36.1	107 26.8	+ 1
536	41 51.9	72 25.9	+ 3	574	45 27.9	107 43.2	−17
537	41 52.5	72 57.5	+43	575	45 19.0	107 54.8	+20
538	41 52.3	72 29.6	+ 1	576	45 18.8	107 21.7	−26
539	41 52.0	72 34.1	+13	577	45 17.9	108 54.2	+22
540	42 05.0	72 34.4	+20	578	45 07.8	107 23.6	−27
541	41 58.2	72 21.0	+15	579	44 58.1	107 20.4	−11
542	42 06.2	72 40.0	+24	580	45 47.5	108 30.3	+ 1
543	42 07.2	72 45.5	+25	581	45 59.5	107 58.8	− 3
544	41 59.2	72 28.1	+ 7	582	46 07.7	107 33.2	+ 7
545	42 11.3	72 51.3	+33	583	46 17.5	107 13.3	+ 9
546	42 09.3	72 22.0	+22	584	46 16.2	106 39.9	+20
547	42 04.6	72 18.5	+31	585	46 16.5	106 26.2	+24
548	42 09.1	72 28.8	+10	586	40 06.7	88 13.6	+ 6



A. Isostatic Anomalies



B. Normal Warp Anomalies



C. Final Warp Anomalies

FIG. 1.—Maps showing isostatic and warp anomalies

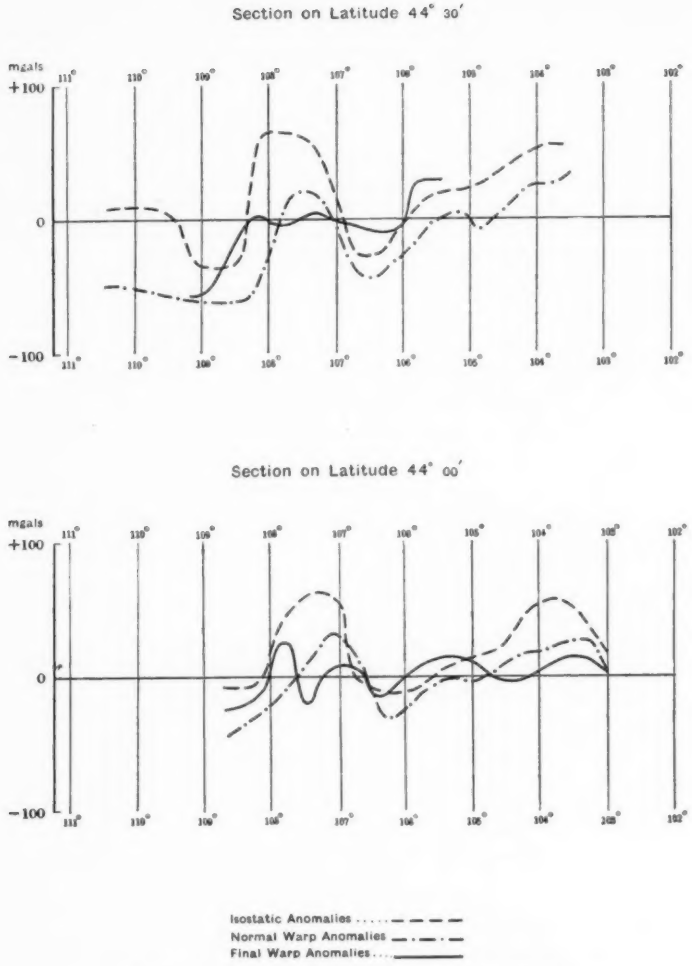


FIG. 2.—Sections showing isostatic and warp anomalies

The superficial rocks overlying a downwarped area of the crust will normally be derived from the sediments laid down in the basin resulting from the downwarp, though subsequent narrowing and compression may have folded and crushed them up into mountains. It follows, therefore, that the negative gravity anomaly due to the crustal downwarp is increased by the low density of the superficial sedimentary rocks. Similarly, in an upwarped area the positive anomaly due to the upwarp is increased by the greater density of the superficial rocks. These in a strong upwarp will normally be igneous rocks of the deeper parts of the granitic layer or else the more ancient, and so more compact, sedimentary rocks, thrust up by the upwarp. Hence the normal warp anomalies in which no geologic or topographic correction for superficial rocks has been included will tend to give a somewhat exaggerated picture of the crustal warping. A general chart of normal warp anomalies should therefore bring out strongly the main tectonic warpings of the crust, though some of the minor details may be lost through the absence of corrections for superficial geology.

A diagram of normal warp anomalies for the whole of the United States is shown in the Figure 3.

Data in some parts, especially in the Middle West, are very meager, and some major warpings may have been missed; but in the east they are sufficient to show very plainly the general structure.

Figure 4 shows a section from the east coast along Lat. 36°. Discussing the Appalachian geosyncline, Pirsson writes: "It should be noted also that during the depression of the geosyncline mountain-making forces were active over the western part of Appalachia, which was then in the condition of a rising geanticline."[6] This geanticline is shown very strongly by the warp anomalies. From about Lat. 31°, Long. 84°, this upwarp runs northeast to Lat. 43°, Long. 31°, or beyond.

The greatest positive anomaly found over this upwarp is +62 mgals.; but anomalies of about +40 mgals. are more usual, corresponding to an upwarp of the crust of about 10,000 feet if normal density (2.67) of superficial rocks is assumed. The Appalachian geo-

[6] L. V. Pirsson and C. Schuchert, *Textbook of Geology* (New York: John Wiley & Sons), Part I, p. 381.

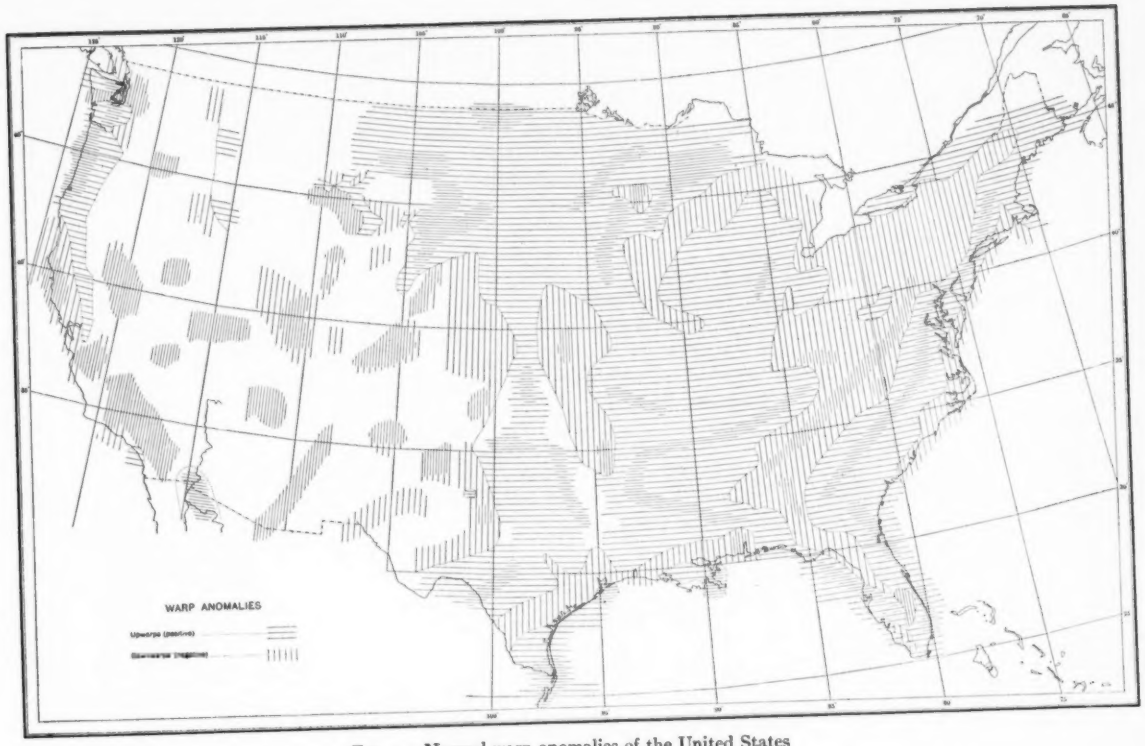


FIG. 3.—Normal warp anomalies of the United States

syncline runs parallel to this upwarp, passing east of the Adirondacks, which appear as a positive mass. The anomalies over the geosyncline are smaller, mostly about −25 mgals., indicating a downwarp of about 7,000 feet; but this may be due to the gravity stations not lying over the main axis of the downwarp shown dotted on the section; the greatest anomaly found (−39 mgals.) corresponds to a depth of about 11,000 feet.

Farther west the Cincinnati geanticline appears as a broad upwarp, but gravity stations are too sparse to define it properly. Dr. Donald Barton has suggested the possibility of an extension of the

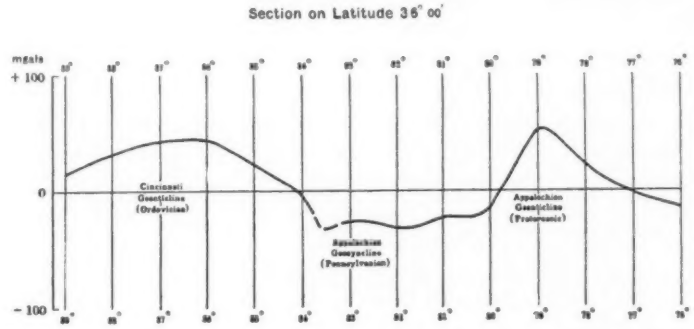


FIG. 4.—Section from the east coast of the United States along latitude 36° showing warp anomalies.

Appalachian geosyncline southwest joining on to an Iberian–North Louisiana geosyncline.[7] The evidence of the warp anomalies is emphatically opposed to this. From about Lat. 33°, Long. 87°, the Appalachian geosyncline appears to run southeast toward Florida, flanked by corresponding trends of the Cincinnati and Appalachian upwarps. In the Iberian region gravity stations are too few, and probably the details as extrapolated by me in the chart are incorrect. There is evidence of a pronounced upwarp trending south-southeast from gravity station No. 301 (+73 mgals.); and east of this at station No. 96 the warp anomaly is −43 mgals., disclosing a downwarp. It is likely that this downwarp continues south, paralleling

[7] "The Iberian Structural Axis," *Jour. Geol.*, Vol. XLI (1933), pp. 225–42.

the upwarp to the west; but if so, the gravity stations have been unfortunately placed. It must be borne in mind that in the southern regions the great depth of unconsolidated sediments will of themselves cause a large negative anomaly; so, if the area is downwarped, very large negative normal warp anomalies should be obtained. The data as they stand, and taking into consideration the necessity for a positive geologic correction, favor the existence of a strong upwarp under the "Iberian Peninsula." More gravity stations are needed in this area, where rapid changes in anomaly may be found between closely spaced stations.

Reference may be made to the gravity station at Seattle, No. 56. The isostatic anomaly here is -94 mgals., but the normal warp anomaly is -74 mgals. Thus the warp anomaly, though large, is not excessive. It would probably be further reduced by geologic correction, but a large negative anomaly would certainly still remain. It has frequently been stated that this anomaly is purely local, but the chart (Fig. 3) shows it to be part of a narrow but extensive downwarp.

Gravity observations at sea in the Gulf of California would be interesting. The isolated land station at the head of the Gulf, No. 65, has a positive warp anomaly $+25$ mgals. If high positive anomalies are obtained in the gulf, this feature would be similar to the Gulf of Cambay in India, which appears to be a rift overlying an upwarp.[8]

The broad area of negative anomalies in the Middle West shown in Figure 3 may be illusory, owing to the chance omission of any gravity observations over upwarps. On the other hand it may, in fact, be a broad downwarp balancing a broad upwarp under Eastern Canada.

The writer hopes that geologists will critically examine these results. If warp anomalies prove satisfactory in areas where the geological structure is well known, then they may help to solve problems where the geological structure is obscure. The station numbers in Table III and IV are those given in the *Principal Facts for Gravity Stations in the United States* issued by the Coast and Goedetic Survey, in which details of the isostatic anomalies, etc., are given.

[8] E. A. Glennie, "Gravity," *loc. cit.*

THE FACTOR OF FATIGUE IN ROCK EXFOLIATION

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ABSTRACT

By analogy with the behavior of crystalline aggregates of the metals, rocks might be expected to fail by fatigue under repeated applications of stress much below the strength as measured in ordinary laboratory tests. To determine whether the temperature stresses set up in a rock by insolation are sufficiently great to cause failure by fatigue, an experiment has been performed in which a coarse-grained granite has been subject to temperature differences of 110° C. for a period corresponding to 244 years of natural insolation. Photomicrographs show no change in the rock.

INTRODUCTION

The importance of temperature change in rocks as a cause of weathering (exfoliation) has long been emphasized by geologists. Many have held that the diurnal changes of temperature from the heat of the sun to the cold of night develop in exposed rocks stresses which are great enough to fracture the rock, causing spalling off of slabs and the shaping of "boulders of disintegration."

Dr. Eliot Blackwelder has ably reviewed the evidence for and against this hypothesis of insolation as an agent of weathering. His conclusion reads as follows:

> To the question whether, assuming that insolation ever causes rock breakage, the process is rapid and is capable of producing its effects in a day, a negative answer seems to be enforced. If insolation can cause fracture at all, it would seem to be only by the accumulation of small increments of strain continued for centuries.
>
> Theoretically, solar heating resulting in expansion should cause exfoliation in dense rocks. Sudden cooling should cause cracks normal to the surface. In granular rocks, the complex alternation of these movements between crystals should cause the mass to disintegrate into rubble or sand. Many writers on this subject ascribe all three of these results to insolation. It does not seem to me that we have good evidence in favor of the first and second of these opinions, but the third—granular disintegration, produced by the accumulation of minute strains over long periods of time—seems better supported.
>
> As yet we do not appear to have satisfactory evidence that insolation is ever the sole unaided cause of rock fracture. However, it seems at least possible that by setting up strains in rocks it aids in loosening the cohesion of the

minerals, facilitates the entrance of moisture, and thus promotes the breakage of the rock by expansion due to chemical change.[1]

In almost all investigations of the effect of thermal changes on rocks the factor of fatigue has been completely neglected. In only one experiment known to the writer has the attempt been made to find the effect of repeated cycles of temperature change. Trowbridge and Wentworth subjected a rock to temperature changes from 0° to 100° C. over a period corresponding to five years, taking each cycle as the equivalent of one day.[2]

In 1932 the writer and Mr. A. J. Holmberg of Ohio State University discussed the probable effect of repeated alternations of temperature. Theoretically it seemed possible that the stresses resulting from differential expansions in the separate crystals and also from the temperature gradient in the rock might cause fracture, if the temperature change were repeated often enough. The reason for supposing that the repetition of cycles causes greater effect than one single cycle is purely the analogy with the work of the metallurgists on fatigue of crystalline aggregates of the metals. These have been found to break under a repeated stress equal to less than half that required to fracture them on the first application of pressure.

Accordingly, to test this factor, Mr. Holmberg and the writer set up an apparatus which automatically subjected a rock specimen to periodical heating and cooling. That apparatus was kept running for a period equal to 58 years. Then the writer moved to Harvard University where a new apparatus was built and the experiment continued, using the same rock specimen. This work was done in the laboratory of mining geology under the direction of Dr. L. C. Graton. The writer takes pleasure in making acknowledgment of the helpful co-operation of Dr. Graton and Mr. E. B. Dane, Jr., and the valuable criticism of Dr. Kirk Bryan.

In all, the specimen has been subjected to temperature changes corresponding to the diurnal changes of 244 years. This paper reports the results of the experiment and discusses the validity of reasoning from these results to conditions of rock weathering.

[1] Eliot Blackwelder, "The Insolation Hypothesis of Rock Weathering," *Amer. Jour. Sci.*, Vol. XXVI (1933), p. 110.

[2] *Ibid.*, p. 108.

FATIGUE

It is a well-known fact that steel structures, when subject to stresses repeated frequently over a long period of time, tend to fail under a stress much lower than their strength when measured in a single laboratory test carried to the point of failure. This phenomenon has been intensively studied by mechanical engineers and metallurgists, and a great deal has been learned about the mechanism of fatigue failure and its relationship to crystal structure. It is interesting to study their work in the light of its possible applications in the field of geology.

Relation between number of cycles and strength.—Of prime interest is the relation between the strength of a specimen and the number of cycles of stress application to which it is subjected before failure occurs. The method of the experiment is obvious. The stress is decided first and then repeated applications of this stress or, in some cases, alternations from plus to minus values of this stress are continued until failure of the specimen occurs. When the results of a series of these experiments were plotted in graph form, we have the relation shown in Figure 1. Here we notice two things of interest: (1) The strength when the stress is applied over a great many cycles is only 75 per cent of the strength when fracture occurs on the first cycle. (2) The relation is of exponential form, so that as the stress is increased above that required to produce fatigue failure after a great many cycles, the number of cycles necessary to cause failure decreases exponentially.

This latter fact is important in the planning of an experiment to duplicate geological conditions. By applying to rock in the laboratory cycles of stress greater than that caused by the diurnal temperature change in nature, the same effect is produced as would result from many times that number of cycles in weathering.

Relation of fatigue failure to crystal plasticity.—The study of fatigue has been confined almost entirely to metal crystals which are much more ductile than most rock crystals. Fatigue seems to occur only after the true elastic limit has been reached, and there is a definite plastic set produced in each cycle of stress. This set may not persist after the external force has been removed, as is the case in general in plastic deformation, but nevertheless it is there,

as shown by the fact that the material does not follow Hooke's law faithfully. The name "elastic hysteresis" is applied to this phenomenon. Now we ordinarily think of rocks as being brittle at atmospheric pressure, having no plastic properties. For this reason a question might be raised as to whether the analogy of fatigue in metal crystals can be applied to rocks. Apparently no measurements of the elastic properties of rocks have been made which are precise enough to show this elastic hysteresis, if it exists. However, it is only

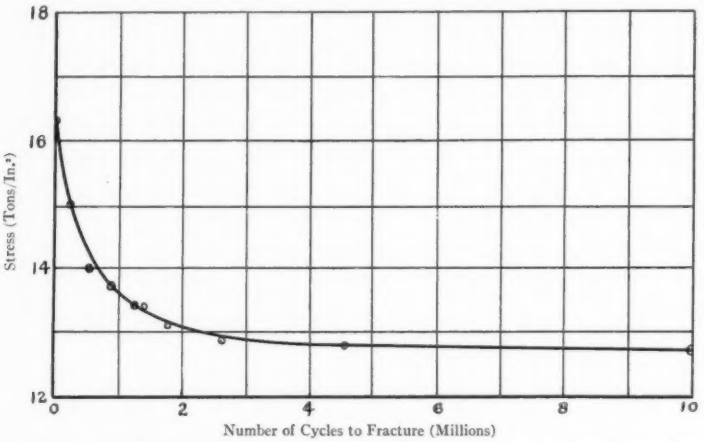


FIG. 1.—Fatigue tests on mild steel (after H. J. Gough, *The Fatigue of Metals* [London: Benn, 1926], p. 48).

recently that methods of measuring which are sensitive enough to show this hysteresis in metals have been developed.

The metallurgists have discovered a fact which ties fatigue failure even more closely to the plasticity of crystals. H. J. Gough and others have shown that fatigue failure occurs along planes of slip or twinning in the crystal, and that this slip or twinning is caused by the applied force. Gough may be quoted on this point: "The initial location of failure is always concerned with plastic movements along certain crystallographic planes. Fatigue failure must, therefore, be considered as a consequence of slip."[3]

[3] "Crystalline Structure in Relation to Failure of Metals," *Amer. Soc. for Testing Materials*, Vol. XXXIII, Part II (1933), p. 95.

There is a further suggestion toward success in tracing the effects of fatigue on rocks. It is often possible to detect the fatigue cracks long before complete failure actually occurs. Photomicrographs of a polished surface of crystalline rock which has been subjected to repeated cycles of stress might be expected to show such evidence in advance of fracture.

Application to rocks of principles illustrated by metal crystals.—If, then, we know fatigue to be a property of metal crystals, may we expect the phenomenon to be exhibited also in rocks? There are two major differences between metal crystals and rock crystals. First, the metal crystals, as already pointed out, are much more ductile. Second, they are much simpler in atomic structure and are held together by different binding forces. Consequently, we might expect to find that somewhat different principles apply to the two types of crystals.

However, it can be shown that rock crystals are plastic when subject to a high confining pressure such as obtains deep in the earth's crust. Also, laboratory work by the writer has shown a number of other direct analogies between the behavior of rocks under high pressure and the behavior of ductile metals. For example, processes familiar in the case of metals and called "creep," "elastic after-working," "work-hardening," and "raising of the elastic limit by cold-working," have been found duplicated in limestone.[4] Hence, there seems to be no fundamental difference in behavior because of the different types of atomic bonding.

With regard to ductility, the writer has found no record of experiments by mechanical engineers which show a definite relation between the lowering of strength by fatigue and the brittleness of the material. Doubtless work has been done which will show this relation, and then by extrapolating, or interpolating as the case may be, we may get a better idea of what to expect in rock crystals. At present, it would seem good policy to consider fatigue in rock crystals at surface pressures as a definite possibility. The present work has supplied no evidence one way or the other.

[4] D. T. Griggs, "Deformation of Rocks under High Pressure," *Jour. Geol.*, Vol. XLIV (1936), p. 541.

THE EXPERIMENT

In the experiment the attempt was made to approximate the conditions of nature but to control all factors except insolation. Heating was accomplished by radiation from a red-hot electric heater; cooling by a stream of dry air, to avoid the effects of moisture; and the temperature range of the experiment (110° C.) was

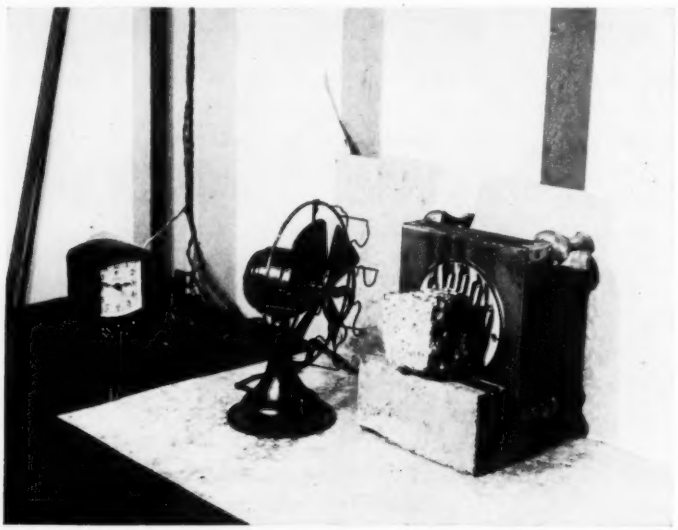


FIG. 2.—Apparatus with granite specimen in place

about as great as is found at any outcrop throughout the year, although greater than the diurnal range of any locality known to the writer.

The polished rock was set on a piece of firebrick so that its face was one inch from the hotplate, as shown in Figure 2. By means of an electric relay connected to the clock, the heater was turned on (with the fan off) for a period of five minutes. Then the heater was turned off and the fan on for a period of ten minutes, when the cycle began again. This apparatus was set up in a basement room where the temperature varies little during the year.

The cyclical change of temperature was determined by the use of a small thermocouple placed against the surface of the rock. The thermocouple was shielded by asbestos so that it was heated only by the rock surface. In order to insure that the rock with which the thermocouple was in contact had been heated to its normal temperature, the whole unit was moved continually over the face which was being heated. Because the area of the unit was small compared

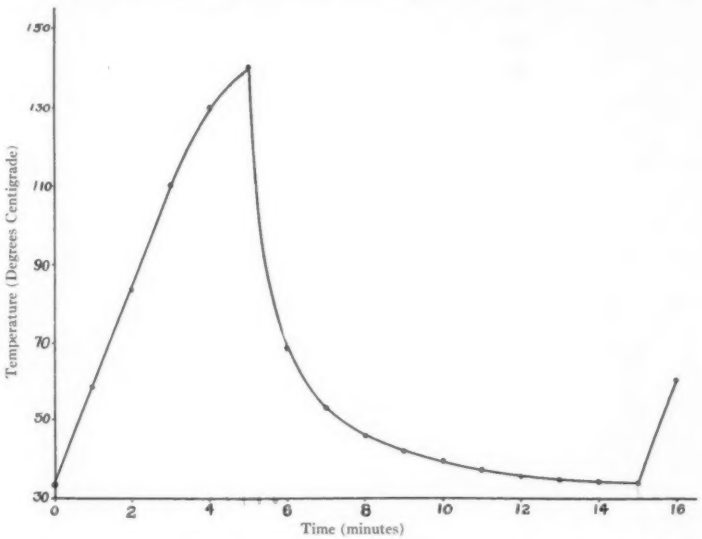


FIG. 3.—Surface temperature of the granite during one cycle of heating and cooling.

to the area of the face, the error occasioned by its shielding effect was negligible, as was shown by check measurements on one cycle in which the thermocouple was inserted only at the end of the heating and cooling periods.

The nature of the heating and cooling curve is shown in Figure 3. The experimental error of any of the measurements is less than 5° C. The values fall on two exponential curves as would be expected from heating and cooling by sources of constant temperature; hence the approximate accuracy seems assured.

From the curve it should be possible to calculate the temperature gradient, the stress distribution, and the progress of the heating wave through the rock. However, the analysis is complicated and the writer has not yet succeeded in obtaining a solution of exact form. An approximation which is probably accurate enough for the geological problem will be discussed later.

The first experiment was begun on April 5, 1932. Four small specimens of granite, syenite, felsite, and sandstone, each about 1.5 inches square, were simultaneously exposed for a period corresponding to 3.8 years, figuring one cycle equal to a day. No change in the rock surface could be discerned except that a few flakes of the polished surface fell away. These had apparently been loosened originally in the process of polishing.

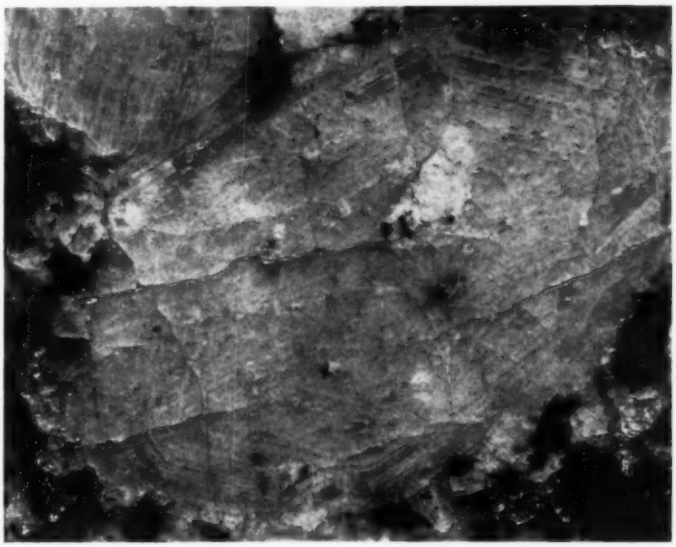
On May 1, 1932, a new specimen of Minnesota pink granite ($3\frac{1}{4}\times3\frac{1}{4}\times2\frac{3}{4}$ inches) was placed in the apparatus and the second experiment begun. This granite was chosen because of its coarseness of grain, since it was believed that the effect of the differential stresses in the several mineral grains due to the various coefficients of thermal expansion in the different crystallographic directions would be of more importance than the spalling stresses due to the temperature gradient set up in the rock.

The specimen was subjected to approximately 89,400 cycles or the equivalent of 244 years of exposure to insolation in nature. The run was almost continuous. Four times the power actuating the relay was shut off and the heater accidentally remained on for from one to four hours. This long exposure heated the specimen to a temperature of near 400° C. It is remarkable that even this great heat caused no change in the specimen.

Results.—Because of the fact that so little effect was shown in the preliminary experiment, it was decided to make photomicrographs of several critical areas on the face of the specimen before it was subjected to the temperature changes and to use these as a standard for comparison during the progress of the experiment. Unfortunately the first series of photographs are not as sharp and clear as the later ones. However, where the earlier photos are sharp, it is possible to trace every feature exactly as it shows in the later series. Consequently, it is believed there was no change during the whole

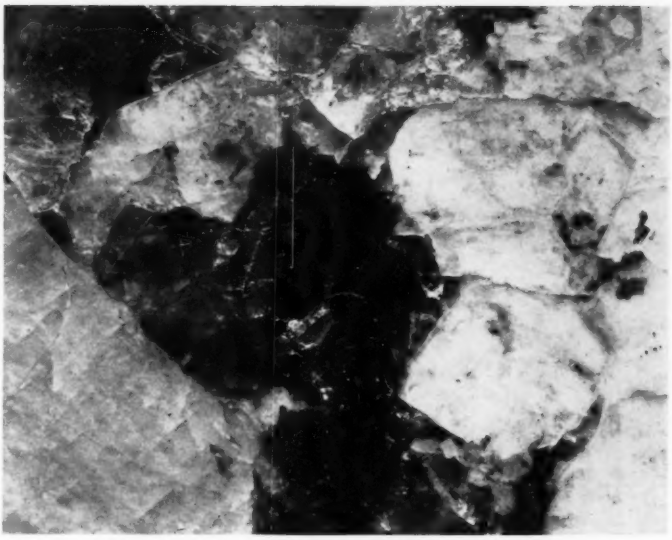
PLATE I

A. Photograph of Large, Zoned Feldspar Crystal Showing Cracks after 58 Years of Experimental Insolation. (These cracks were all present in the specimen before subject to temperature change. ×4)



B. Same Area after 244 Years of Experimental Insolation. (×4)

PLATE II



A. Photograph of Quartz-Feldspar Boundary after 58 Years of Experimental Insolation. (×9)

B. Same Area after 244 Years of Experimental Insolation. (×9)

experiment. The series of good photomicrographs cover a period of 186 years, or the last 76 per cent of the duration of the experiment.

The position of the areas photographed is shown on the full-size picture of the specimen (Fig. 4). These areas were chosen to show all types of contacts between mineral grains. Also, one was taken in the middle of the big feldspar crystal where a number of small

FIG. 4.—Full-size photograph of granite specimen

cracks terminated. The lighting was arranged with great care to bring out to the best advantage the cracks in the minerals, since it was thought that these would be the most sensitive detectors of stress in the specimen. Some of the photomicrographs are reproduced in Plates I and II. They show no change in detail so far as can be seen in these reproductions. Further, a very close study of enlargements has shown absolutely no difference in the regions of the photographs where the lighting was identical. The excellence of these photographs is entirely due to the help of Mr. E. B. Dane, Jr., who kindly co-operated in this work.

It can definitely be said that the 89,400 cycles, or the equivalent of 244 years of heating and cooling with a temperature range of 110° C. or 200° F., produced on this coarse-grained granite no effect which could be detected under moderate microscopic magnification.

APPLICATION TO GEOLOGY

Blackwelder pointed out that two types of failure might result from the temperature changes of insolation, namely, exfoliation and disintegration of the mass due to the differential coefficients of expansion of the crystals themselves. The former will be considered first.

Before the results of this experiment may be applied to exfoliation, we must consider its difference in conditions from those which apply in nature. Two such differences are at once apparent. First, the specimen is smaller than the rock faces which are believed to exfoliate by insolation. Second, the rate of heating and cooling is much more rapid than that which occurs in nature. These conditions must be examined to see how they affect the stresses set up in the rock.

It might seem natural to assume that size determines whether a specimen will exfoliate or not. Engineers find that beyond a certain limit of size concrete blocks in a pavement will buckle.

In the case of the pavement, cohesion between the concrete and the ground is small, and it takes little force to pull it loose and bulge it up in a buckle. Conditions in the experiment and in massive rocks in nature are very different. The thickness of the rock is great compared to the depth of penetration of the cyclical heat-wave. Hence the surface layer is prevented from expanding laterally by the rigidity of the rock underneath. In this case, the shear stress is proportional to the temperature gradient and is independent of the size of the specimen. If, however, the specimen is so thin that the rigidity of the unaffected part is not sufficient to overcome the shear stress set up by the temperature changes, then the specimen has a tendency to bend and become convex toward the heat source. This bending is in response to the shear stress and hence decreases that stress in the rock considerably below the value it would have if no bending took place.

An exact solution of the problem of stress distribution experimentally induced within the rock is very desirable for comparison with that set up by insolation in nature. Because of the complex boundary conditions and the complex form of the heating curve, this has not yet been feasible. It is to be hoped that some mathematician may succeed in this problem so that we may know the exact magnitude of stresses induced by insolation.

Although the exact solution could not be reached, an approximation may be obtained by assuming the heating curve to be of sinusoidal form and using the equations for an infinitely thick body. The second assumption introduces small error because of the small amplitude of the temperature variation at the back face of the specimen. The application to nature is quite justified because the heating and cooling curves are sensibly sinusoidal and the extent of the rock is so great as to be equivalent to infinite thickness in many cases.

With these assumptions, the following equation for the range of the temperature variation may be used:[5] Range $=\theta_0 e^{-\frac{x}{h}\sqrt{\frac{\pi}{T}}}$, where T = period = 15 min. = 900 sec.; θ_0 = temperature variation at the surface = 2 (average temperature − minimum temperature) = 65° C.; h^2 = diffusivity = .0155 for granite; and x = depth in specimen (centimeter).

In the case of the experiment:

$$R_1 = 65e^{-\frac{x}{\sqrt{.0155}}\sqrt{\frac{\pi}{900}}} = 65e^{-.477x}.$$

For a depth of 1 cm., $R_1 = 40.4°$ C. For a depth of 7 cm., $R_1 = 2.3°$ C. (temperature range at back face of rock). Range = 10 per cent of surface variation when $R_1 = 11°$ C., $110°/11.0° = e^{+.477x}$, $x = 3.73$ cm.

For diurnal temperature change, $R_2 = 60e^{-.0488x}$, where $\theta = 60°$ C., $T = 86{,}400$ sec., and $h^2 = .0155$. For a depth of 10 cm., $R_2 = 36.8°$. For a depth of 1 m., $R_2 = .46°$. Range = 10 per cent of the surface variation, $R_2 = 6°$, $60/6 = e^{+.0488x}$, $x = 47.2$ cm.

[5] Ingersoll and Zobel, *Mathematical Theory of Heat Conduction* (Ginn and Co., 1913), p. 34.

Also, the temperature gradient may be obtained from this equation:[6]

$$\text{Temp. grad.} = \frac{\partial\theta}{\partial x} = -\left(\frac{1}{2h}\sqrt{\frac{\pi}{T}}\right)\theta_0 e^{-\frac{\pi}{h}\sqrt{\frac{\pi}{T}}}\left\{\sin\left(wt - \frac{x}{h}\sqrt{\frac{\pi}{T}}\right) + \cos\left(wt - \frac{x}{h}\sqrt{\frac{\pi}{T}}\right)\right\}.$$

In the experiment:

$$\frac{\partial\theta_1}{\partial x} = 26.2e^{-.477x}(\cos .477x - \sin .477x),$$

where $t = \text{time} = T/2$, arbitrarily chosen. For $x = .01$ cm., $\partial\theta_1/\partial x = 25.9°/\text{cm.} + 5°$ (steady state gradient) $= 30.9°/\text{cm.}$ For $x = 1.0$ cm., $\partial\theta_1/\partial x = 7.0° + 5° = 12.0°/\text{cm.}$ For $x = 7.0$ cm., $\partial\theta_1/\partial x = .73° + 5° = 5.73°/\text{cm.}$

For diurnal change in granite:

$$\frac{\partial\theta_2}{\partial x} = 1.465\, e^{-.0488x}(\cos .0488x - \sin .0488x),$$

for $x = .01$ cm., $\partial\theta_2/\partial x = 1.465°/\text{cm.} = \text{maximum.}$

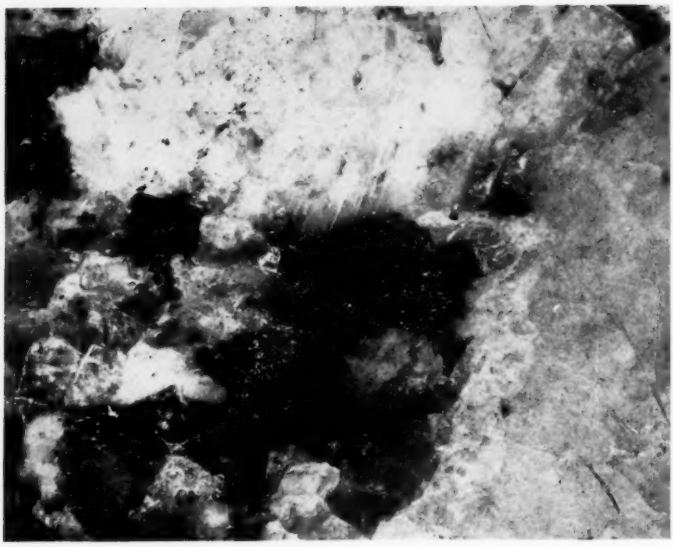
$$\left.\frac{\partial\theta_1}{\partial x}\right|_{max} \Big/ \left.\frac{\partial\theta_2}{\partial x}\right|_{max} = \frac{30.9}{1.465} = 21.1.$$

These computations show, first, that the penetration of the cyclically applied heat into the specimen is practically restricted to the outer 3.7 cm., which is just half the thickness of the specimen. Further, the temperature range drops off rapidly near the surface. Now, since shear stress is proportional to the temperature gradient, we get some idea of the relative stresses developed in the specimen and in nature. It appears that the stresses in the granite block during the experiment are 21.1 times as great as in an outcrop of similar rock where the diurnal variation is 60° C. or 108° F.

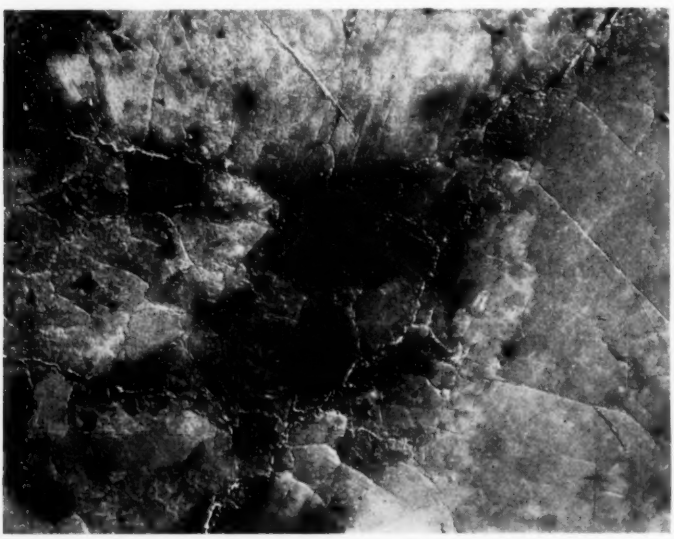
In the case of disintegration of the mass due to the thermal expansion of the crystals, there is an altogether different type of stress distribution. Here the differential stress is caused merely by the variation in coefficients of expansion in the different mineral species,

[6] Ingersoll and Zobel, *op. cit.*, p. 36.

PLATE III



A. Photograph of Quartz, Feldspar, and Hornblende Crystals after 244 Years of Experimental Insolation Which Produced No Change in This Area. (×9)



B. Same Area after Ten Days (Equivalent to $2\frac{1}{2}$ Years) of Temperature Change under Humid Conditions. (Note etching of cracks and pitting of surface)

and in different directions in the same crystal. The intensity of the stress is here controlled by size of crystals and by maximum temperature difference during the cycle.

In the granite chosen for the experiment the grain is coarse, average diameters being nearly half an inch. The temperature range at the surface was 110° C. which is probably greater than is ever found on rock outcrops. Therefore, stresses set up in the surface layer of the granite were probably greater than those caused by insolation at rock outcrops. No granulation or failure resulted.

PLANS FOR FURTHER EXPERIMENTATION

Since it has been shown that temperature changes alone are not effective in producing rock exfoliation or disintegration in the laboratory, it is now planned to continue experiments under different conditions in an attempt to evaluate quantitatively the importance of the several agents of rock weathering. The next factor to be investigated is the effect of humid conditions.

An experiment is in progress in which the heating is accomplished in the same manner as in the former experiment, and its intensity is the same. The cooling, however, is effected by a fine spray of water. Although this is a more efficient cooling agent than a stream of dry air, it is not thought that the temperature falls much more rapidly than in the case of air cooling. Figure 3 shows that air cooling is very rapid in the first stages.

This experiment thus retains very nearly the same stress conditions due to temperature change, but introduces the chemical effect of hydration. Also, since the water used is tap water, it is in equilibrium with the atmosphere and consequently contains some CO_2 and oxygen in addition to other impurities, which will presumably cause additional chemical changes in the rock.

It is yet too early to give any detailed description of the results of this experiment. It seems appropriate, however, to state that the first ten days of its operation produced notable changes in the granite block which survived three years of the former experiment without any change at all. Cracks of the exfoliation type opened up around the edges of the specimen; it lost its polish; and the cracks observed in the photomicrographs widened and deepened. Plate III shows the effect produced on one of the areas photographed in

the first experiment which showed no change throughout the duration of that experiment, but was considerably etched and pitted during ten days of this experiment under humid conditions.

SUMMARY

In an experiment designed to test the effectiveness of fatigue in rock exfoliation by insolation, a specimen of granite was subjected to artificial temperature changes in a fifteen-minute cycle varying from 32° C. to 142° C. Other factors such as hydration, etc., were eliminated or reduced to a minimum by using radiant heat, dry cooling, and maintaining the temperature above freezing. This experiment has been continued for 89,400 cycles, or the equivalent of 244 years of diurnal temperature change, if one cycle is considered equivalent to one day.

Because of the facts that (1) the temperature range is greater than in nature; (2) the temperature gradient is much steeper; and (3) the crystal size is large, the stresses set up are much larger than would be produced by insolation at a rock outcrop. It appears that the shear stresses of the exfoliating type set up in this experiment are about twenty-one times as great as those produced in granite outcrops when subjected to a diurnal temperature change of 60° C. By analogy with fatigue curves of metals, this "overstressing" should produce on the rock the effect of a period much longer than that of the experiment. Therefore, it is suggested that this experiment actually corresponds to the effect on a rock of insolation over somewhat more than a thousand years.

Photomicrographs of the surface before and after the experiment show no change. The small cracks in the component crystals and along their boundaries have the same extent after the experiment that they did before. It seems that the effect of temperature changes over a thousand years is not sufficient to cause any exfoliation or disintegration of granite.

It is thought by many geologists that hydration is a more powerful agent of disintegration than pure temperature change. In an attempt to verify this experimentally, the writer is conducting a new experiment where the effect of moisture will be superimposed on the temperature change. In the first ten days (equivalent to two and a half years) a marked change was produced in the specimen.

THE CLINTON IN NEW YORK

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ABSTRACT

Change of facies has resulted in the naming of more formations than exist in the Silurian of New York. The Rochester, Williamson, Sodus, and Reynales formations, or their representatives, have considerable areal extent. The Brewerton and Schroeppel do not appear to be valid units. The Rochester and Irondequoit are here grouped with the Lockport in the Niagara group. The section at Vanhornsville is briefly described and discussed.

INTRODUCTION

The varying character of the lower Clinton sediments of western New York has been previously considered.[1] The present writing deals with the remainder of the Clinton of this state, together with some of the overlying formations, and discusses the problems of correlation eastward to Vanhornsville (Fig. 1), beyond which there are no known exposures of the Clinton. Although Vanuxem[2] has mentioned "some loose masses and fragments" south of Saltspringville and the writer has seen indications of the group there, he has not found it exposed. Some doubt exists concerning portions of the Silurian section east of Clinton, but the present status of the correlation is presented together with some observations on the misunderstood Vanhornsville sequence.

Serious errors of interpretation have resulted from the paucity of exposures in the vicinity of Oneida Lake and the marked change in facies which takes place in this region. The gap has now been bridged by the series of diamond-drill cores obtained by Newland and Hartnagel for the New York State Museum, but difficulties have arisen from attempts to determine the stratigraphy from published thicknesses of the rock units. Through the courtesy of Mr. C. A. Hartnagel the writer has had the privilege of examining the cores, as a result of which the stratigraphy seems somewhat less complex. The num-

[1] J. T. Sanford, "The Clinton in Western New York," *Jour. Geol.*, Vol. XLIII (1935), p. 169.

[2] L. Vanuxem, "Survey of the Third Geological District," *Geology of New York*, Part III (1842), p. 80.

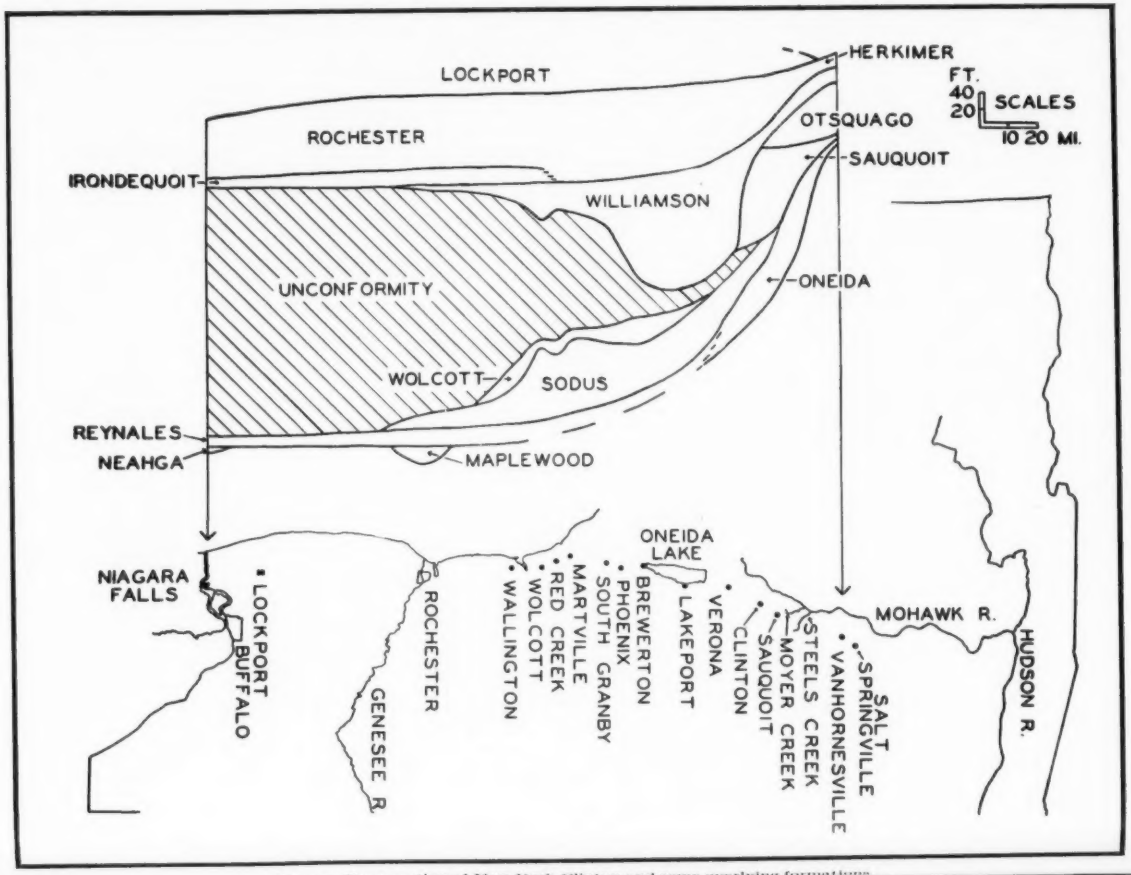

FIG. 1.—Cross section of New York Clinton and some overlying formations

ber of described formations is believed to exceed the number which actually exists. The Brewerton and Schroeppel shales which have been considered to lie above the Williamson are really portions of the Sodus and Williamson, respectively. The Reynales can be traced eastward to the vicinity of Verona, the sandy Martville facies and phosphatic materials being conspicuous. Its exact relation to the Oneida cannot be seen, but it appears to merge into this formation, which, like the Reynales, becomes progressively younger eastward, rising as high as and possibly into the Middle Clinton. The Sodus is found as far east as Clinton village, and a fauna indicative of Williamson age has been discovered at Vanhornsville. The Wolcott limestone with associated hematite appears at least as far eastward as Lakeport, although at Lakeport there are indications that it lies below the top of the Sodus. The Verona ore is probably a member of the Williamson. The Irondequoit, which is not definitely known to the east of the Oswego quadrangle, is closely associated with the Rochester and may continue eastward in the basal beds of that formation. The Rochester is thought to grade into the Herkimer, which extends to Vanhornsville. The Middle Clinton appears in the eastern sections and is represented by the Sauquoit beds and possibly the Otsquago.

The Clinton is considered to extent downward to the top of the Thorold and upward to the top of the Williamson, the Irondequoit and Rochester being excluded from it and grouped with the overlying Lockport. The present grouping of the formations is indicated in Table I.

THE REYNALES FORMATION

The Reynales may be followed in the drill cores as far east as Verona. The sandy facies (Martville member) is prominent. It consists of sandstones, some of which are limy, with interbedded shales and beds rich in hematite. These are characterized by phosphatic materials, either black sands or nodules. At Verona, which is the farthest well to the east, black sands predominate. These suggest the black sands of the Sauquoit Valley, which occur not far above the Oneida conglomerate, and which lie in the *Mastigobolbina lata* zone of Middle Clinton age. However, the sandstones at Verona are

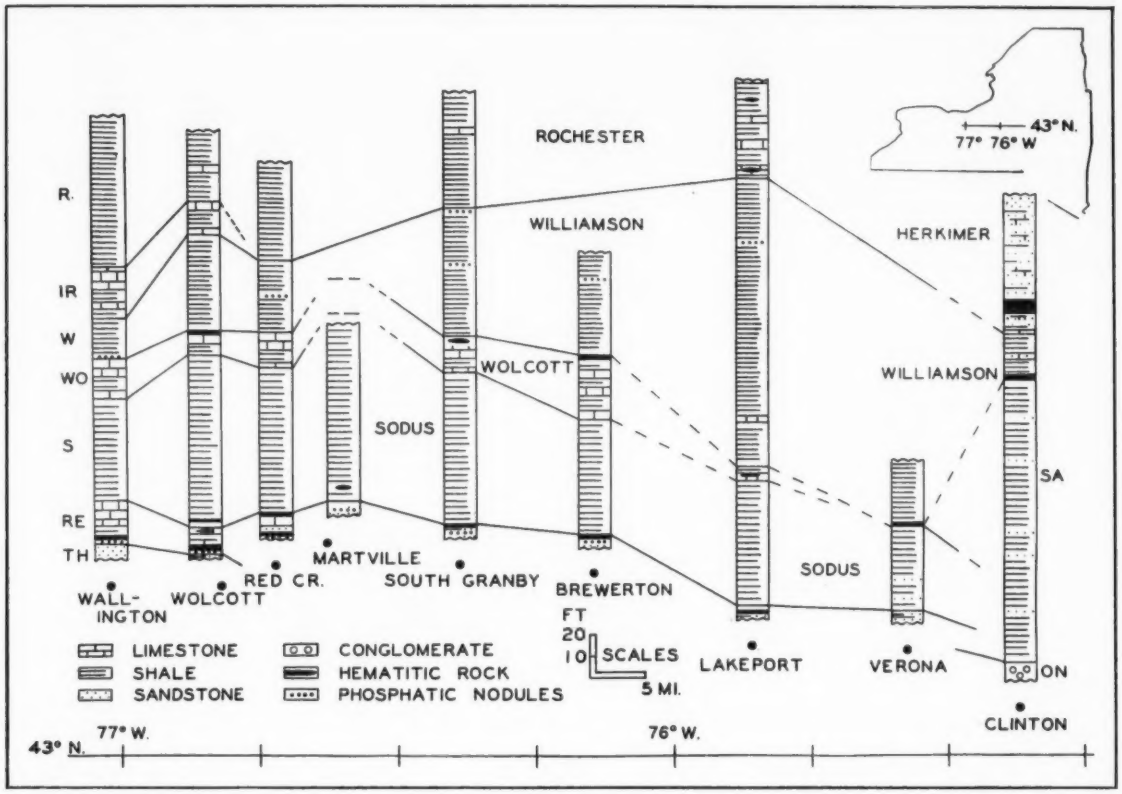

FIG. 2.—Correlation from Wallington to Clinton. The hematite beds are mostly not to scale, and limestone and sandstone symbols in the shales do not indicate the absolute size or number of these rocks. Abbreviations are as follows: *R*, Rochester; *IR*, Irondequoit; *W*, Williamson; *WO*, Wolcott; *S*, Sodus; *RE*, Reynales; *SA*, Sauquoit; *ON*, Oneida; and *TH*, Thorold.

overlain by the Sodus, and hence are retained in the Reynales formation.

The correlation of this easternmost phase of the Reynales is based chiefly on stratigraphic position. The phosphatic materials, which have been noted from Niagara to Verona and in the somewhat younger basal Clinton farther east, cannot be considered as demonstrating contemporaneity, but do suggest that the various facies of the Reynales have some genetic elements in common. The Reynales

TABLE I

Niagara Group	Lockport
	Rochester—Herkimer
	Irondequoit
Clinton Group	Williamson
	Otsquago
	Sauquoit
	Oneida
	Wolcott
	Sodus
	Reynales
	Maplewood
	Neahga
Medina Group	Albion
	Cataract
	Queenston

probably rises in the section as it is traced eastward. This tendency has been demonstrated[3] to the west where fossils occur.

THE SODUS FORMATION

The purple shales characteristic of this formation at Sodus Bay thicken eastward, and green beds become more prominent. At South Granby there are 72 feet of the formation and even more at Martville where the total thickness is unknown, but at Brewerton and Lakeport it has thinned to between 50 and 60 feet. It is represented at Clinton, but is probably missing a few miles to the east in the Sauquoit Valley, for at that locality, from beds which are probably not far above the Oneida conglomerate, fossils of the *Mastigobolbina lata*

[3] Sanford, *op. cit.*, pp. 177–80.

zone have been collected. At Clinton this portion of the section is not well exposed, but from green shaly beds to the south of Clinton Village the writer has collected specimens of *Coelospira nitens* (Vanuxem),[4] which is a characteristic Sodus fossil. The Sodus must wedge out very rapidly to the east of Verona.

The Sodus, according to the present interpretation, includes the shales excavated from the canal at Brewerton and designated by Chadwick the Brewerton shale.[5] In support of this he has cited the fauna obtained by Smith,[6] which he stated is composed mostly of Rochester species.

The writer visited the Brewerton locality in 1933 and obtained a few specimens from loose material along the canal bank. More recently he has examined the collections at Syracuse University and believes that the fauna represents a fairly typical Sodus association. Additional information concerning the Ostracoda, obtained since the publication of Chadwick's paper, has considerably clarified Silurian stratigraphy. The ostracod material from Brewerton is rather poor, but is referred to *Zygobolba rectangula* Ulrich and Bassler. It is certainly a Sodus form. The presence of *Eophacops trisulcatus* (Hall) among the fossils collected by the writer is also indicative of Sodus age.

It is the writer's belief that the remainder of the fauna is more typically Sodus than an examination of the older faunal data might indicate. There seems to be no place in the section at Brewerton and Lakeport for a Brewerton formation, as the portion of the stratigraphic column formerly thought to be included in that formation is occupied by the Williamson.

THE WOLCOTT FORMATION

The Wolcott formation is in some respects similar to the Reynales, which was originally confused with it. It is composed of interbedded limestones and shales with hematite and contains a form of *Pentamerus*. The Wolcott Furnace ore which occurs in the vicinity of the

[4] Professor Chadwick has kindly pointed out the correct name for this fossil referred to as *C. hemispherica* (Sowerby) in my previous paper on the Cinton.

[5] G. H. Chadwick, "Stratigraphy of the New York Clinton," *Geol. Soc. Amer. Bull. 29* (1918), p. 349.

[6] B. Smith, "Notes on the Fossils of the Paleozoic Formation within the Syracuse Quadrangle," *N.Y. State Mus. Bull. 171* (July, 1914), pp. 57–58.

type locality is included, chiefly on the basis of lithology. Chadwick[7] eliminated the confusion regarding the Pentamerus beds, and suggested the correlation of the Wolcott with the beds above the oolitic ore at Clinton. Ulrich and Bassler[8] have demonstrated the younger age of these latter strata.

In this region thin beds of hematite or traces of this mineral are commonly associated with the type of lithology typical of the Wolcott, and probably do not maintain constant horizons. The Wolcott is no exception to this generalization. No hematite beds are to be seen in the Wolcott in the core from Red Creek, but at South Granby there is about a foot of limestone at the top of the formation which contains traces of this mineral. At Brewerton there are traces of hematite in a coarsely crystalline limestone, and less than an inch of hematite somewhat higher, which, with the intervening beds, might be included in the formation. In the Lakeport core the Wolcott type of lithology is still present, but here it is overlain by a few feet of shale of a Sodus facies. This shale contains an unidentified ostracod which bears a close resemblance to a Sodus form. The question arises as to whether a fauna may not sometimes shift upward as a shoreline is approached. What seems more likely is that the lithology has shifted downward and that the Wolcott represents a regressive tendency. The Wolcott facies of the Lakeport well, with the $3\frac{1}{2}$ feet of overlying shale, is provisionally included in the Wolcott.

THE ONEIDA FORMATION

It is not strange that the age relationships of the Oneida conglomerate have been variously interpreted, for, like the Reynales, it is not a time unit. The conception of a range in age due to shifting shorelines was emphasized by Grabau,[9] although the true age relationships were more obscure at that time than they are at present. Hartnagel[10] cited the position of the Medina below the Clinton and

[7] *Op. cit.*, p. 344.

[8] E. O. Ulrich and R. S. Bassler, "Paleozoic Ostracoda: Their Morphology, Classification and Occurrence," *Md. Geol. Surv., Silurian* (1923), p. 338.

[9] A. W. Grabau, "Guide to the Geology and Paleontology of the Schoharie Valley in Eastern New York," *N.Y. State Mus. Bull. 92* (1906), pp. 123–24.

[10] C. A. Hartnagel, "Stratigraphic Relations of the Oneida Conglomerate," *N.Y. State Mus. Bull. 107* (1907), pp. 33 and 36.

on this basis, and reports of *Arthrophycus*, made the Oneida the equivalent of the Medina. The finding of a Middle Clinton ostracod zone at New Hartford caused Ulrich and Bassler[11] to believe that the Oneida is of Middle or Lower Clinton age. This is correct within a limited area, but as the Oneida is traced westward it becomes progressively older. At New Hartford (see discussion of Sauquoit beds) it is overlain by the Sauquoit, and at Clinton by the Sodus formation. If, as seems probable, the sandy beds at the bottom of the Verona core represent the Martville member of the Reynales, then this still older formation separates the Oneida from the Sodus at this locality. The Oneida and Reynales, with the associated Maplewood and Neahga, represent an initial phase of the eastwardly transgressing Clinton sea. The Oneida ranges from Reynales up to the time of the *Mastigobolbina lata* zone in age.

THE SAUQUOIT BEDS

The term "Sauquoit" was proposed by Chadwick[12] to include "all the shale and sandstone beds between the Oneida conglomerate and the oolitic ore bed in the Oriskany and Sauquoit valleys." It is well that he designated "Swift Creek, north of Sauquoit village," and the vicinity as the type locality, for the finding of *Coelospira nitens* (Vanuxem) south of Clinton indicates that the Sodus formation is present just above the Oneida conglomerate there. Near New Hartford, Ulrich and Bassler[13] recognized the *Mastigobolbina lata* zone and the same zone was found at Willowvale by the writer in beds that must occur not far above the Oneida conglomerate. It appears that the Sauquoit beds do not include strata of Lower Clinton age at the type locality, and therefore should be limited to beds above the Sodus at Clinton. The exact position of the boundary is not known. As they are overlain by the *Mastigobolbina typus* zone it will not be possible to determine whether or not the upper Sauquoit beds are of Middle or Upper Clinton age until more paleontological evidence is available.

The stratigraphic position of the 5 feet of gray shale overlying the Oneida formation at Vanhornsville places it somewhere in the

[11] *Op. cit.*, pp. 337 and 346.

[12] *Op. cit.*, p. 341. [13] *Op. cit.*, p. 374.

Sauquoit. The unit pinches out rapidly to the west of Clinton, and is apparently absent in the well cores, but again the paleontological data are insufficient.

THE OTSQUAGO FORMATION

The Otsquago sandstone, which is 60 feet thick at Vanhornsville, thins rapidly to the west and is but a few feet thick at Moyer's Creek. It is not present at Clinton. Red and cross-bedded to an extreme degree, it exhibits a shoreward facies which would probably become increasingly important eastward were the Clinton better represented in that direction.

The exact age of the Otsquago formation has not been determined. Chadwick[14] thought it might be equivalent to his Bear Creek, which was of Lower Clinton age. Its position above the Sauquoit beds and Oneida conglomerate and below the Williamson formation would make it fall somewhere between the *Mastigobolbina lata* and *Mastigobolbina typus* zones. It does not seem to correlate with any formation except possibly a portion of the Sauquoit beds.

THE WILLIAMSON FORMATION

The Williamson formation can be traced from Rochester, where it is represented by 6 feet of dark shales, to Vanhornsville, where *Bonnemaia oblonga* Ulrich and Bassler, a representative of the *Mastigobolbina typus* zone to which the Williamson belongs, occurs in a thin lens of hematite-rich rock just beneath the Herkimer sandstone. Its greatest thickness of 137 feet appears in the diamond-drill core from Lakeport. The character of the Williamson at Clinton has been adequately described by Ulrich and Bassler.[15] Like the underlying formations, it bears witness to a shoreline and source of sediment to the east. In Williamson time the shoreline had migrated as far to the east as it was to move during this transgression. The western exposures of the Williamson formation are characterized by thin black shales rich in the remains of a planktonic fauna (*Monograptus clintonensis* [Hall]), but in which benthic forms are also represented. Eastward toward the supply of sediment it thickens, green beds become more prominent, and the benthic fauna becomes

[14] *Op. cit.*, p. 343.

[15] *Op. cit.*, pp. 344–45.

more numerous. As the former shoreline is still more closely approached, the Williamson formation again becomes thinner. Phosphatic nodules appear at at least one fairly constant horizon and seem to check the other stratigraphic data. Their significance remains uncertain, but they probably represent an accumulation of phosphatic material over a long period of time or else re-working. Their position at the base of the Williamson in the Wallington section might strengthen either interpretation. The most definite horizon occurs within the formation, and what little faunal evidence is at present available from the cores indicates that the rocks above and below are paleontologically similar. It may be that there is a physical break within the Williamson formation.

The Williamson as here defined includes the beds at Phoenix designated by Chadwick[16] as the Schroeppel shale. In support of this view he lists a fauna from the type locality which has Rochester affinities, and which includes *Plectambonites transversalis* (Wahlenberg) (probably a *Sowerbyella*). Specimens collected by the writer, and others examined in the collection at Syracuse University, indicate that the *Sowerbyella* (*Plectambonites*) present is not *transversalis*, the Rochester species, but is a form characteristic of the Williamson formation elsewhere. Reaffirming the Williamson age of these rocks is the presence of *Monograptus clintonensis* (Hall) and *Retiolites geinitzianus venosus* (Hall). There are a number of forms represented which suggest the Rochester shale, but it becomes apparent that there may be some overlapping of Rochester and Williamson forms in the thicker sections. The beds at Phoenix may represent a somewhat younger phase of the Williamson which is not seen in the western exposures where the formation is thinner.

The strata which have been assigned to the "Schroeppel" include portions of both the Rochester and the Williamson formations in the South Granby well, and in the Lakeport core they are predominantly of Rochester age.

The hematite bed in the vicinity of Verona was designated the Verona iron ore by Chadwick[17] and correlated by him with the oolitic ore of the Clinton region. He also suggested that it grades

[16] *Op. cit.*, pp. 351 and 352. [17] *Ibid.*, p. 346.

into limestone at Lakeport. It was not known at that time that this limestone is of Lower Clinton age, whereas the oolitic ore (Williamson) is Upper Clinton. The correlation with the oolitic seems more logical as this is the only other well-defined hematite bed occurring in this portion of the stratigraphic column in this vicinity. Unfortunately the scanty data from the Verona core help but little, and a visit to the old ore pits northwest of Verona, which are undoubtedly those described by Vanuxem as west of Verona, was disappointing as the ore was not available. The fossils listed by Chadwick are mostly from Clinton, and to my knowledge none is from Verona.

This interpretation would necessitate a rather important disconformity at the base of the Verona ore in the Verona district, and would include the Verona in the overlying shales as a member of the Williamson.

THE IRONDEQUOIT FORMATION

The Irondequoit formation is typically a fossiliferous, coarsely crystalline limestone and, although shale beds become more prominent in the lower part in the vicinity of the type locality and eastward, the limestone is characteristic. It may be seen in the outcrop from localities in Canada to the eastern part of the Oswego quadrangle in New York State. The typical lithology is missing in the Red Creek well, which is somewhat to the west of the easternmost exposure. East of the Oswego quadrangle the Irondequoit is probably represented by shales, but similarity to the Rochester makes a separation impractical at present. In the Lakeport core there are several feet of limestone and hematite at the base of the Rochester which may correspond to the Irondequoit. It was correlated with the beds underlying the red flux (Kirkland) ore at Clinton by Ulrich and Bassler,[18] chiefly on the basis of stratigraphic position.

The Irondequoit resembles the Rochester and certain facies of the Lockport more closely than it does the underlying formations. A survey of the principal papers dealing with the fauna shows that of sixty-two species listed from the Irondequoit formation (excluding the reefs) twenty-eight also occur in the Rochester shale, and this list will undoubtedly be extended. Very few Irondequoit species are

[18] *Op. cit.*, p. 385.

common in the underlying formations. Of the species common to the Rochester and Irondequoit formations, a fair number occur frequently in both. It seems that the Irondequoit should be grouped with the Rochester and Lockport and it is therefore excluded from the Clinton group and placed in the Niagara group.

THE ROCHESTER FORMATION

The Rochester formation, including the Herkimer sandstone, which was correlated with it by Ulrich and Bassler,[19] but which Chadwick[20] correlated with beds somewhat lower, can be traced from Vanhornsville westward to the Niagara region and over into Ontario, Canada. At Vanhornsville the Herkimer member is represented by 12 feet of sandstone, whose base is associated with the Williamson horizon (see discussion of the section at Vanhornsville) and which thickens rapidly westward, being well over 70 feet on Steele's Creek, according to Vanuxem.[21] At Clinton the Rochester is represented by the Kirkland (red flux) ore bed[22] and the overlying rocks. These rocks are typical of neither the Herkimer sandstone nor the gray shales seen farther west. They are composed of interbedded sandstones and shales with limy sandy beds marked with traces of hematite and giant ripples. These beds probably represent deposition influenced by a shoreline which lay not far to the east. The Rochester horizon was not encountered in the Verona well as it started too low in the section, but exposures in the old ore pits contain fossiliferous limestones and sandstones. The core from Lakeport contains about 86 feet of interbedded shales and sandstones with limy hematitic beds at this horizon. The character of this material indicates a transition between the strata previously described and what has been considered the more typical Rochester facies to the west. It includes horizons ascribed to the Lakeport and Schroeppel by Chadwick,[23] who carried the Rochester formation at Lakeport up into what the writer believes to be Lockport. The Schroeppel shale has already been discussed. There is at present little evidence to necessitate the use of the term Lakeport, although with a

[19] *Ibid.*, pp. 346 and 348.

[20] *Op. cit.*, p. 351.

[21] *Op. cit.*, p. 82; see also p. 257.

[22] Ulrich and Bassler, *op. cit.*, p. 348.

[23] *Op. cit.*, pp. 350 and 353.

more detailed study of the well cores it may be possible to distinguish members within some of the present formations. At South Granby, about 27 miles to the west of Lakeport, there are 54 feet of shales at the top of the core, which are more like the limy shales of the sections to the west. About 85 feet of the Rochester formation occur in the Genesee Gorge at Rochester, including the Gates member, and nearly 70 feet in the Niagara Gorge. It continues to thin westward in Ontario.

The off-shore phases of the Rochester contain extremely fossiliferous beds from which well over two hundred species are noted in the literature. Faunal relationships with the Irondequoit have been discussed. It has at least thirty-two species noted by various authors as occurring in the Lockport formation, and faunal dissimilarity with the formations below the Irondequoit is marked. Although there are over one hundred and fifty forms listed from the Lockport, the formation is sparingly fossiliferous when considered as a whole, the specimens are not always well preserved, and some of the species listed are not common throughout the area. It is the writer's belief that future study will show even greater faunal similarity between the Rochester and the Lockport. It must also be remembered in reviewing the relationships of the Irondequoit, Rochester, and Lockport formations that faunal similarity is obvious in the common forms seen in the field and that slight differences in species may represent evolutionary changes which took place throughout the deposition of the three units.

The Herkimer sandstone is united with the Rochester formation, first, because of the gradation of the better-known shaly into the sandy Herkimer facies; second, because of the stratigraphic position of the Herkimer and its proximity to the Williamson horizon; third, because what little faunal evidence there is, although Chadwick[24] interpreted it otherwise, seems to point in this direction. It has already been indicated that the beds at Phoenix, with which Chadwick correlated the Herkimer member, contain a strong assemblage of Rochester forms which are rather misleading. From a sandstone not far above the Otsquago at Moyer's Creek the writer collected *Parmorthis elegantula* (Dalman), which is a common Rochester spe-

[24] *Ibid.*, p. 351.

cies, although not limited to that formation. The Rochester is an example of change in facies within a time unit, owing to geographical variation at a time of widespread marine transgression in New York State. The easternmost Herkimer may represent incipient regression, which tendency was continued and increased in Lockport time, as indicated by the intraformational structures and character of that formation.

LIMITS OF THE CLINTON

The term "Clinton" has been used in several ways, and hence it becomes necessary to redefine the group in accordance with the rules of stratigraphic nomenclature in current use.[25] The early use of the term was somewhat ambiguous, since the details of correlation were but partially understood at that time. Vanuxem's report[26] on the Third District and Hall's[27] on the Fourth indicate their intention to exclude the Thorold and the Oneida from the base of the Clinton and the Rochester from its top (Table II), but Vanuxem states that the Oneida is a part of the Clinton group although separated from it. Formations in western New York, at least as old as the Oneida conglomerate, were included in this group, and the Oneida to the east of Clinton is as young as the beds above this formation at this locality. It is not possible to determine definitely at what point Vanuxem placed the upper limit of the Clinton group at Clinton. He undoubtedly included beds of Rochester age, but he did not know that these strata correlated with the Rochester, and made it very clear that he placed the Rochester in the overlying Niagara group. There are thus contradictions in the original definition of the Clinton.

When genetic relationships are considered it is possible to form groups not greatly at variance with the intentions of Hall and Vanuxem. The formations fall into three divisions: the first, in

[25] *Geol. Soc. Amer. Bull. 44* (1933) Part II, pp. 436 and 437. In this report on "Classification and Nomenclature of Rock Units," it states, in art. 13, that "groups shall receive geographic names, which shall be subject to the same restrictions and regulations as apply to formation names," and in art. 10, "The redefinition of a unit to change its content of rocks, where such change is demanded by later work, does not necessarily render renaming advisable."

[26] *Op. cit.*, pp. 79–80.

[27] James Hall, "Survey of the Fourth Geological District," *Geology of New York*, pp. 58–79.

which a terrestrial influence is strongly felt, includes the Queenston, Cataract, and Albion; the second, which is marked by variability, is composed of formations lying between the top of the Thorold and the top of the Williamson; and the third, which is made up of the Irondequoit, Rochester, and Lockport, is dominated by a marine

TABLE II

NOMENCLATURE OF THE CLINTON SECTION

Vanuxem*	Chadwick,† 1917	Ulrich and Bassler,‡ 1923	Sanford, 1936
	Correlated with Phoenix (Schroeppel) shale and Herkimer sandstone	Herkimer sandstone	Rochester formation (Herkimer member)
	Kirkland iron ore	Red flux—Rochester	(Kirkland member)
		Keefer-Irondequoit	?
	Vanhornsville sandstone	Brewerton Williamson	Williamson formation
CLINTON GROUP	Oolitic ore Correlated with Verona ore	Oolitic ore	Oolitic ore (Verona)
	Sauquoit beds Sodus member Martville member	Middle 130 feet Clinton	Sauquoit beds
	Maplewood? member		Sodus formation
ONEIDA CONGLOMERATE	Oneida conglomerate	Oneida conglomerate	Oneida formation

* Vanuxem, *op. cit.*, pp. 75, 76, and 80.
† Chadwick, *op. cit.*, Fig. 2, p. 333.
‡ Ulrich and Bassler, *op. cit.*, pp. 346 and 347.

influence. The three groups taken together form a normal sequence of uplift followed by transgression and regression with what is probably a wedge of terrestrial influence (the Otsquago), thickening shoreward.[28] It is proposed to include these formations in New York State in three groups as indicated in Table I.

[28] The writer does not intend to imply that the Otsquago is necessarily a terrestrial deposit.

The term "Clinton" offers certain difficulties, but it is so firmly imbedded in the literature that it seems best to retain it. The base of the section at Clinton is not complete, there being no representative of the basal shales which occur in western New York, but it is nearly so if the Oneida is considered to replace the Reynales, which it does from a genetic point of view. As has been previously pointed out, the Reynales and Oneida become progressively younger from west to east and hence the base of the group varies in age from place to place. The Middle Clinton, which is absent in western New York, is represented at Clinton by the undifferentiated Sauquoit beds. The upper limit is as yet undetermined definitely at Clinton, but lies somewhere between the two ore beds. When more fossils are found in this part of the section, this difficulty may be eliminated. Many of these uncertainties center about the position of the type locality in a shoreward facies, and the resulting lack of fossils and diversity of sediments. The more normal facies farther west might make a much better measuring stick, even though the Middle Clinton is lacking. It seems impractical in dealing with a group as varied as the Clinton to consider but one locality for purposes of definition, but rather to deal with a region. The writer does not expect that the natural grouping of the formations of New York State will necessarily apply to other regions, but this is immaterial as long as interregional correlations are not based on group definitions.

THE VANHORNSVILLE SECTION

Much of the misunderstanding concerning the section at Vanhornsville has arisen from the inadequacy of the available descriptions. A brief summary of the formations is therefore presented in Table III.

This section emphasizes the loss of strata from both the top and the bottom of the Medina-Clinton-Niagara sequence as it is traced eastward. The Lockport (and also the overlying Vernon) has disappeared from the top and the Rochester is represented by but 12 feet of sandstone, the Herkimer, which from a time point of view may range as low as the Irondequoit or even the Williamson. The lithology of the upper part of the *Mastigobolbina typus* zone (Williamson) just below it is so similar that the two might be rather close-

ly related. On the other hand, the lithology is definitely Herkimer, and the sequence is similar to that farther west. It seems probable that the Herkimer here represents incipient regression, and under these circumstances it might be expected to be slightly older than the greater extent of the formation.

TABLE III

FORMATIONS OF THE VANHORNSVILLE SECTION

	Feet
Camillus formation	
Slabby, buff shales (only the basal beds present)	10
Rochester formation	
Herkimer member; gray sandstones with shale partings	12
Williamson formation	
Intermingled sandstones and shales containing a thin limy hematitic bed (about 0.1 ft.), somewhat variable and containing *Bonnemaia oblonga* Ulrich and Bassler, a representative of the Mastigobolbina typus zone	2
Shale, with a zone of phosphatic nodules	16
Otsquago formation	
Coarse red sandstones with which are associated a small amount of fine-grained, smooth shales; the sandstones are extremely cross-bedded	60
Sauquoit beds?	
Gray shale	5
Oneida formation	
Conglomeratic and sandy beds with black phosphatic sands and black nodules	8
Frankfort formation (Ordovician)	
Dark-gray shaly rocks	

The 16 feet of shale below the *Mastigobolbina typus* zone should probably be included in it, as a rather indifferent specimen found in it is referred to *Bonnemaia oblonga* Ulrich and Bassler. Near its top is a zone of phosphatic nodules associated with a few phosphatized fossils, frequently broken, and this evidence, although far from conclusive, strengthens the argument.

The Otsquago formation has been previously considered. There

is at present little to establish the identity of the Sauquoit beds except their position above the Oneida conglomerate. The Oneida has probably continued to become younger eastward and may very well be of Middle Clinton age.

CONCLUSIONS

The stratigraphy of the Clinton and adjoining formations of New York is perhaps less complicated than has been supposed. Change of facies has diversified the appearance of the formations, some of which do not necessarily represent time units.

The Rochester formation is correlated with the Herkimer sandstone. The typical Irondequoit is missing in the east, but the formation is probably represented in the basal Rochester beds. The Williamson can be traced from Rochester, where *Monograptus clintonensis* (Hall) characterizes the fauna, to Vanhornsville, where *Bonnemaia oblonga* Ulrich and Bassler is a prominent fossil. The Wolcott, with included hematite, appears to be older at Lakeport than it is farther west. It may represent a regressive tendency. The Sauquoit beds, which are at least in part of Middle Clinton age, are best represented in the vicinity of Clinton. The presence of *Coelospira nitens* (Vanuxem) indicates the occurrence of the Sodus at Clinton. Its extent there is unknown. The Reynales is represented at Martville and eastward chiefly by the sandy Martville facies with the associated phosphatic and hematitic materials. It probably continues to rise in the section, as has been demonstrated farther west. The Oneida, which is the basal Clinton formation in the eastern sections, likewise becomes younger in this direction.

THE FLYNN CREEK DISTURBANCE, JACKSON COUNTY, TENNESSEE[1]

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ABSTRACT

A small, intensely disturbed area in Jackson County, Tennessee, has been mapped and studied in detail. All the data accumulated indicate a crypto-volcanic origin of the structure. A magnetometer survey showed a well-developed magnetic high offset 4 miles to the south-southwest. The history of this disturbed area is interpreted as follows: Shortly preceding Chattanooga deposition an explosion took place near the surface, blowing out a crater 2 miles in diameter and 300 feet deep. The Ordovician limestones forming the floor and walls of this crater were shattered into breccia composed of angular fragments of varying sizes imbedded in a matrix of smaller fragments and "rock flour." The deeper parts of the crater were filled with redeposited breccia, either as talus breccia or as bedded breccia deposited in a fresh-water lake that occupied the crater at one time. The Chattanooga sea invaded central Tennessee and filled the crater with black mud, now represented by about 250 feet of black shale. Fort Payne chert was later deposited upon the relatively smooth surface of the black shale.

INTRODUCTION

In September, 1935, the writers studied 6 square miles of highly disturbed beds along Flynn Creek about 5 miles south of Gainesboro, Jackson County, in the southwestern part of the Gainesboro

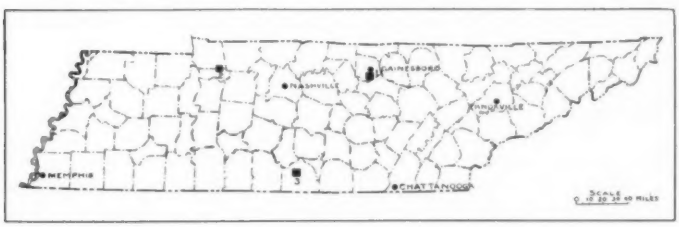

FIG. 1.—Index map of Tennessee, showing the location of (1) Flynn Creek disturbance, (2) Wells Creek basin, and (3) Howell structure.

quadrangle (Fig. 1). This area is readily accessible from the Gainesboro–Double Springs highway (State Highway 56), which crosses

[1] Published with the permission of the State Geologist, Tennessee Division of Geology.

the northeastern corner of the map (Fig. 2). Roads along the valleys of Flynn Creek and Cub Hollow permit easy access to the localities of interest. The intensely disturbed area is about 2 miles in diameter and has its center about ½ mile northwest of Antioch School and the small settlement of Clenny.

Previous Work

Safford[2] was the first to note the abnormal structural features of the beds in this part of Jackson County. He pointed out that

> another area of disturbance is in the upper part of the valley of Flynn's Creek, in Jackson County. This area is limited in extent, and has comparatively little importance; yet the formations are greatly disturbed. The rocks are seen to dip at high angles, and are occasionally almost vertical. The valley is narrow, and the hills on each side high. In their normal position the *siliceous* is at the top of the series of formations, and the *Black Shale* next below. In several places both are brought down, by great folds and faults, to the bottom of the valley, and, at one point, may be seen abutting against the Nashville Formation. One fault shows a displacement of a thousand feet. The lines of disturbance run nearly north and south.

In 1926 Lusk[3] mapped the Gainesboro quadrangle for the Tennessee Division of Geology. Lusk interpreted the disturbed area in the vicinity of Clenny as a pre-Chattanooga sinkhole and in 1927 published a brief account of this feature.[4]

Bassler[5] visited the locality and noted that "on three sides of this area the dips suggest that the region may be another example of crypto-volcanic structure such as that described by Bucher for Adams County, Ohio."

Geography

The Gainesboro quadrangle is part of the highly dissected eastern Highland Rim region of Tennessee, which has an average elevation of 1,000 feet above sea-level. In the southwestern part of the quad-

[2] J. M. Safford, *Geology of Tennessee* (Nashville: S. C. Mercer, 1869), p. 148.

[3] R. G. Lusk, "Geology and Oil and Gas Resources of the Gainesboro Quadrangle, Tennessee," *Tenn. Div. Geol. Bull. 45* (1935).

[4] R. G. Lusk, "A Pre-Chattanooga Sink-Hole (Gainesboro Quadrangle, Tennessee)" *Science*, N.S., Vol. LXV (1927), pp. 579–80.

[5] R. S. Bassler, "Stratigraphy of the Central Basin of Tennessee," *Tenn. Div. Geol. Bull. 38* (1932), p. 143.

rangle the Cumberland River and its larger tributaries have cut down through the resistant cherty limestone capping the rim to the general level of the Central Basin of middle Tennessee, which is about 550 feet above sea-level. The writers were unable to detect any control of the local drainage pattern by the local structural features.

Stratigraphy

The consolidated rocks exposed in the area range in age from Ordovician to Mississippian. The formations shown on the map (Fig. 2) are briefly described, as follows:

Ordovician System

Mohawkian Series

Black River Group

Lowville limestone.—The oldest rocks exposed in the area mapped are thinly bedded, grayish-white to dove-colored, dense limestones of Lowville age. These are exposed on the northeast and north sides of the hill along Flynn Creek northwest of Antioch School. This outcrop, which was the only Lowville noted, occupies the center of the disturbed area and is composed of many large, disconnected blocks, some of which are several acres in size. Lusk recognized that some of the limestones in this outcrop were different from other limestones exposed in the Gainesboro quadrangle, but did not commit himself as to their age. Since this limestone is relatively unfossiliferous, the formation was only tentatively correlated in the field with the Lowville of the Central Basin on the basis of its faunal content and lithology. This correlation was later confirmed by R. S. Bassler, who kindly identified the following characteristic Lowville bryozoans:

Rhinidictya nicholsoni Ulrich
Graptodictya proavia Eichwald
Escharopora subramosa Ulrich
Orbignyella wetherbyi Ulrich
Phyllodictya frondosa Ulrich

The true thickness of the Lowville was not determined because of the highly disturbed condition of the beds, but it appears likely that not less than 50 feet of the formation is exposed.

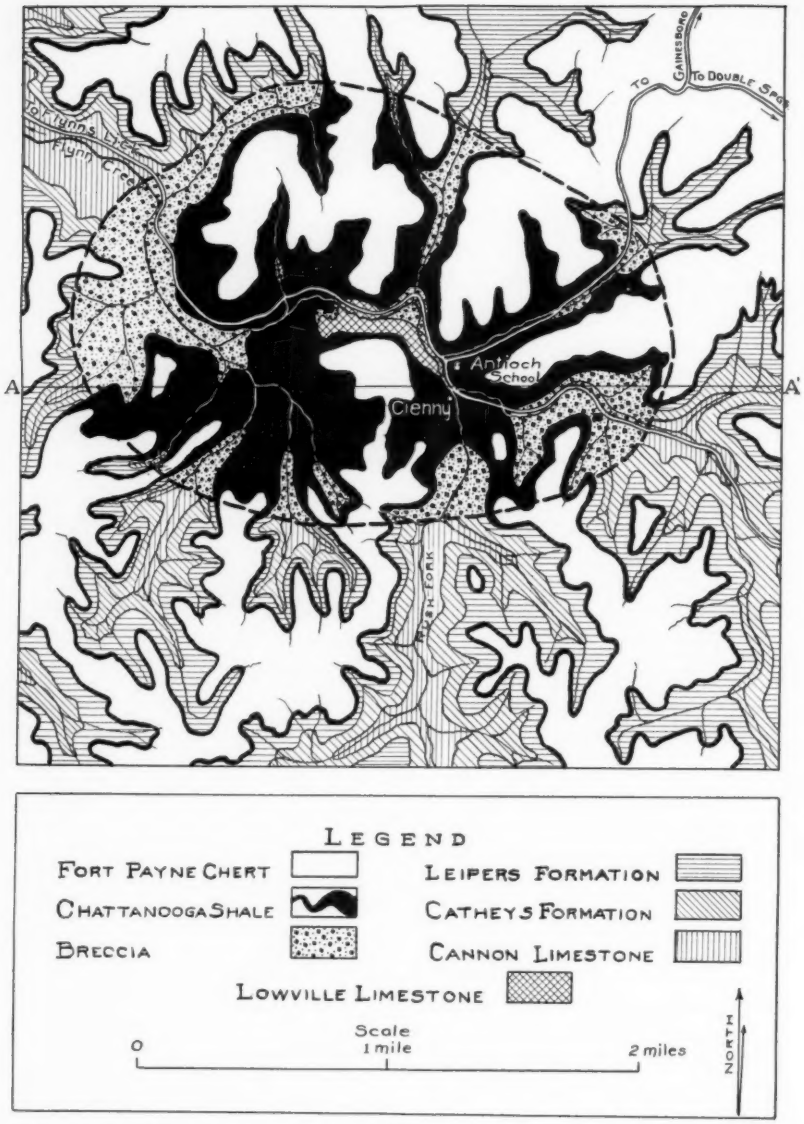

FIG. 2.—Areal geologic map of the Flynn Creek area. The heavy broken line indicates the approximate border of the circular area of intense brecciation.

TRENTON GROUP

Hermitage formation.—In the normal stratigraphic succession along the eastern edge of the Central Basin, the Hermitage formation overlies the Lowville limestone; but this formation was not found in the Flynn Creek area. It is believed that the Hermitage formation was originally deposited in this region but that during the local deformation and subsequent erosion all traces of it were removed.

Cannon limestone.—The Cannon limestone normally occupies the valleys of Flynn Creek and its major tributaries around the periphery of the intensely disturbed circular area shown in Figure 2. Also, isolated blocks of Cannon limestone were identified within the disturbed area. The formation is a fine-grained, bluish dove-colored, massive to thinly bedded limestone. The best exposures of the Cannon in the area mapped are along Flynn Creek about 2 miles northwest of Antioch School and along Rush Fork about ½ mile south of Clenny. Lusk[6] reports a thickness of about 200 feet for the Cannon in adjacent parts of the quadrangle.

Catheys formation.—The Catheys formation normally overlies the Cannon limestone in the valleys of Flynn Creek and its tributaries around the disturbed area, and is known also in isolated blocks within the disturbed area. The limestone of the Catheys formation is so lithologically similar to the Cannon limestone that they may be separated satisfactorily only on the basis of fossils. In the Flynn Creek area the Catheys is an even-bedded, bluish gray, dense to crystalline limestone with shaly horizons. Locally, the formation is distinctly argillaceous. Cross-bedding was noted at several horizons. Owing to the highly disturbed condition of these beds in the mapped area, it was not possible to determine accurately the thickness of the formation; but there is probably as much as 100 feet present.

CINCINNATIAN SERIES

MAYSVILLE GROUP

Leipers formation.—Overlying the Catheys is the Leipers formation, which consists of thinly bedded, gray to blue, shaly limestone with a thickness of about 100 feet in the vicinity of Gainesboro.

[6] Lusk, "Geology and Oil and Gas Resources of the Gainsboro Quadrangle, Tennessee," *loc. cit.*, p. 24.

The areal distribution of the Leipers formation is practically the same as that of the underlying Catheys.

Breccia

Breccia composed of angular fragments of limestone is exposed at many localities within a circular area of about 4 square miles along Flynn Creek and its tributaries (Fig. 2). The fragments range from the size of peas up to large blocks many feet in dimension and occur in a matrix of powdered limestone. A study of a thin section of this matrix shows highly brecciated, extremely angular fragments of calcite with some chert in a matrix of pulverized, recrystallized limestone. There is some evidence of chertification of the calcite fragments.

This breccia contains blocks and fragments of limestone from the Lowville, Cannon, Catheys, and Leipers formations, and is therefore younger than the Leipers (Maysville). As it is overlain by normally bedded Chattanooga shale, its age must be post-Leipers, pre-Chattanooga.

The types of breccia exposed in this area are: (1) shatter breccia, (2) injected powder breccia, (3) talus breccia, and (4) bedded breccia. These are described in the chronological order of their origin relative to each other.

Shatter breccia.—This widely exposed type of breccia consists of limestone blocks and fragments of various sizes held in place by a matrix of smaller fragments. These fragments are angular to subangular, the only rounding having been by mutual attrition at the time of their explosive origin. It is this breccia that forms the matrix in which are imbedded the large blocks of limestone in the central part of the area. This type is well exposed in the bed of the creek between $\frac{1}{4}$ and $\frac{1}{2}$ mile east of Antioch School.

Injected powder breccia.—This breccia is composed of powdered limestone that was injected dikelike along fractures in the large blocks of limestone in the central part of the area. Stringers, or veins, of this breccia, which range in width from a feather edge to a foot, extend across the blocks, dividing and converging in the manner of forceful magmatic intrusions. This type of breccia is best exposed in a small ravine on the steep-sided hill southwest of Flynn

Creek, 0.1 mile northwest of Antioch School. A study of a thin section of this breccia shows an actual "injected" relationship between a very finely granular, essentially recrystallized mass of calcite and clayey material and the much more coarsely crystalline calcite of the "country rock." Excellent examples of "inclusions" of the coarsely crystalline limestone occur in the "vein" material. The contact between the two is generally quite sharp. Some parts of the slide seem to indicate injection of the finely pulverized material into minor fissures on a microscopic scale. Apparently this took place while the material had a "mushlike" consistency.

Talus breccia.—Exposed at several localities is a breccia composed of subangular blocks and fragments suggesting slight rounding, such as would result from traveling a short distance down a steep slope under the influence of gravity rolling or slope wash. The best exposure of the talus breccia is along the road 0.6 mile east of Antioch School.

Bedded breccia.—The two most available exposures of this interesting type of breccia are on the right side of the road 0.2 mile northwest of Antioch School and on the hillside rising behind the house south of Flynn Creek about 0.7 mile northwest of Antioch School. The thickness of these bedded deposits varies up to a measured maximum of 12 feet. The lower beds, which cover an irregular tangle of blocks of limestone (either shatter or talus breccia), contain more or less rounded fragments as large as 4 inches. The fragments in this bedded deposit grade in size from coarse grained in the lowest beds to medium grained in intermediate beds, and to fine grained in the upper beds of each local sequence. Stratification of the breccia parallels that of the overlying Chattanooga shale.

The most plausible explanation of the origin of this breccia is that it was deposited in a fresh-water lake occupying the depression that existed in the Flynn Creek area for part of the post-Leipers, pre-Chattanooga interval. The uniform stratification, locally suggesting lamination, demonstrates its aquatic origin. It is believed that any Silurian or Devonian epi-continental sea which might have reached this region would have filled the crater with sediments that would have been preserved, for the later Chattanooga sea filled the depression with its sediments and these have been preserved. For this

reason the origin of the bedded breccia is attributed to deposition in a fresh-water lake, such as would have formed in the depression.

MISSISSIPPIAN SYSTEM

Kinderhook Series

Chattanooga shale.—One of the most striking features of the Flynn Creek area is the abnormal thickness of the Chattanooga shale. In the Gainesboro quadrangle, as in the greater part of central Tennessee, the Chattanooga shale is characterized by uniformity in thickness. In the southwestern part of the Gainesboro quadrangle the black shale is usually about 20 feet in thickness. However, in the Flynn Creek area the thickness is greatly increased, and near the junction of Rush Fork and Flynn Creek a continuous section of 133 feet of black shale was measured. This section extends from the level of Flynn Creek up the large hill just north of J. W. Chaffin's store at Clenny. A water well at this store was dug 49 feet below creek level and remained in the black shale, indicating a total thickness of more than 182 feet at this locality. At several places north and west of Clenny continuous exposures of 100–200 feet of black shale were noted.

The Chattanooga shale in this area has its characteristic lithology, being a black, fissile, highly carbonaceous shale. It is overlain by a foot or two of the Maury shale member. Along the creek east of Antioch School the lower 6 inches of the black shale contain more or less rounded fragments of the underlying breccia. Thin sections made from specimens of this basal member show a dense to finely crystalline, dark calcareous rock with considerable pyrite and organic structures (spores and bryozoans [?]). Fragments of calcite, highly colored chert, and glassy quartz are common.

Osage Series

Fort Payne chert.—Overlying the Chattanooga shale on the higher hills in the area is the gray, cherty limestone of the Fort Payne formation. It is possible that at least a part of the New Providence formation is locally present. These lower Mississippian formations are not strongly tilted. The Fort Payne is the youngest formation exposed in the area.

STRUCTURE

REGIONAL SETTING

In the Gainesboro guadrangle, as throughout the eastern Highland Rim, the strata are essentially horizontal; and dips greater than 5° are rare. In such a region, characterized by relatively undeformed strata, the presence of a small area of highly disturbed, contorted, and brecciated strata, locally vertical and overturned, is of more than passing interest.

Only two other areas of intense deformation are known in central Tennessee. One is in the Wells Creek basin in southeastern Stewart County, which has recently been described by Bucher.[7] The other is located just north of the village of Howell in the north-central part of Lincoln County. The present writers expect to do more detailed field work on the Howell structure (Fig. 1).

STRUCTURAL FEATURES OF THE FLYNN CREEK DISTURBANCE

The structural features of this area may best be described in two parts: (1) the underlying, intensely deformed strata; and (2) the overlying, relatively undeformed strata. The first series includes strata of the Lowville, Cannon, Catheys, and Leipers formations; the second, the Chattanooga shale and the Fort Payne chert. These two series of strata are separated by a marked unconformity with maximum differential relief of approximately 300 feet within ½ mile (Figs. 3 and 5). The "plane" of the unconformity coincides with the pre-Chattanooga surface, which was a closed topographic basin about 2 miles in diameter and about 300 feet below the level of the surrounding area. In the center of this depression was a hill, composed chiefly of large blocks of Lowville and possibly older limestone, that rose 200 feet above the general floor level.

UNDERLYING, INTENSELY DEFORMED STRATA

The intensely deformed strata of Lowville, Cannon, Catheys, and Leipers age are limited to a circular area with a diameter of about 2 miles in the center of the area mapped. The minor details of structure within this small area could not be determined, as these forma-

[7] W. H. Bucher, "Cryptovolcanic Structures in the United States," *XVI Intl. Geol. Cong. Rept.* (1933). Preprint, Dec., 1935, pp. 12–16.

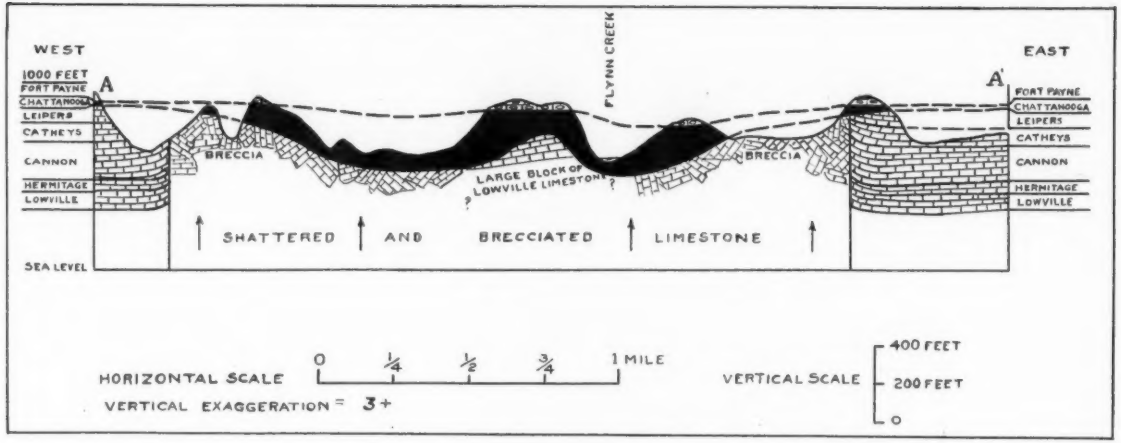

FIG. 3.—East-west structure cross section along the line AA' on Figure 2

tions occur in blocks, varying in size from several acres down to small fragments, and abutting against each other at all possible variations of strike and dip. Many of the contacts between the blocks are obviously fault contacts; but these are too numerous, too closely spaced, and too irregular to be mapped on any practical scale.

The large blocks of limestone occupying the central, circular area are apparently imbedded in a matrix of shatter breccia composed of small fragments of limestone in a groundmass of powdered limestone. The injected powder breccia has flowed around the blocks and has

FIG. 4.—Photograph taken 0.8 mile southeast of Antioch school, showing the dip into the central uplift on the south flank and the thrusting-away from the center of uplift on that side.

forced its way into fractures in the limestone in a method similar to dike intrusion and flowage. Certain beds in the blocks themselves, and also in the nearby strata, disturbed but not disconnected from the country rock, are also brecciated.

In general, the major structural feature consists of a circular uplift which has raised a small central mass of blocks of Lowville limestone vertically into juxtaposition with the Leipers formation, a vertical distance of approximately 500 feet. Surrounding the small central uplift of Lowville limestone is a ring of breccia containing blocks of all the Ordovician formations included in the disturbance. In the outer portion of this ring the less disrupted blocks of Leipers,

Catheys, and Cannon limestone join with the strata of their respective formations *in situ*. On the eastern, northern, and western flanks of the structure the strata dip at varying angles away from the central uplift. On the southern flank the strata dip in toward the center of the uplift. This latter attitude is explained as the result of thrusting outward from the center, evidence for which is seen in an exposure on the south bank of Flynn Creek about 0.8 mile southeast of Antioch School (Fig. 4). Here the outward-shoved strata are overturned and thrust away from the center of uplift.

Overlying, Relatively Undeformed Strata

Overlying and covering much of the intensely deformed Ordovician strata is the Chattanooga shale. This shale was deposited upon the topographic surface shown in Figure 5. As shown in the isopach map (Fig. 6), the shale has its normal thickness of 20 feet around the margin of the deformed area, while within the area of intense deformation the shale attains a thickness of more than 250 feet. Such variations in thickness indicate that the black shale filled a pre-existing topographic basin approximately 300 feet deep. Structure contours drawn on top of the black shale show a closed, synclinal basin (Fig. 7).

The overlying Fort Payne chert was deposited upon a relatively level surface of black shale, and hence shows no anomalies of thickness. The local structure of this formation is a gentle dip toward the center of the closed, synclinal basin, paralleling the slightly greater dips of the underlying black shale.

ORIGIN

Any theory of origin for the structural features in the Flynn Creek area must explain the following: (1) a central uplift of approximately 500 feet, bringing relatively old beds (Lowville) up to the level of younger beds (Leipers); (2) the intense brecciation of the Ordovician limestone, and the grinding, or pulverizing, of much of the limestone into "rock flour"; (3) the striking ability of breccia to actually force its way into fractures in unbrecciated limestone in a way suggesting dike intrusion; (4) the shattering of the Ordovician limestone into large blocks, and the irregular jumbling of these blocks; (5) the dip away from the central uplift on the northern, eastern, and western flanks; (6) the dip into the central uplift on the southern flank, and

the thrusting-away from the center of uplift on that side; (7) a closed, irregular topographic depression with 300-feet relief on the pre-Chattanooga surface, the deformation being post-Leipers, pre-Chattanooga in age; (8) the abnormal thickness of the black shale

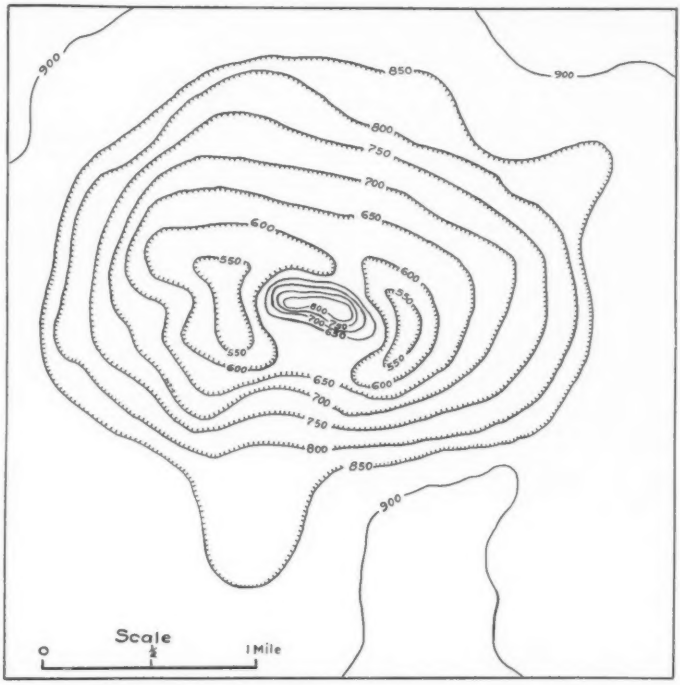

FIG. 5.—Contour map on the pre-Chattanooga topographic surface. Contour interval is 50 feet.

(250 feet); (9) the closed synclinal basin in the black shale and overlying Fort Payne chert, centered over what was originally an uplift (this is rather unusual in a region where anticlines and synclines were formed early in the Paleozoic, and all subsequent diastrophic movements rejuvenated these earlier structures as anticlines and synclines, respectively[8]); (10) a well-developed magnetic high cen-

[8] C. W. Wilson, Jr., "The Pre-Chattanooga Development of the Nashville Dome," *Jour. Geol.*, Vol. XLIII (1935), pp. 471–73.

tered about 4 miles south-southwest of the disturbed area. (This magnetic high is believed to be the surface expression of the postulated buried plug of igneous material responsible for the Flynn Creek disturbance. The offset of 4 miles to the south-southwest is

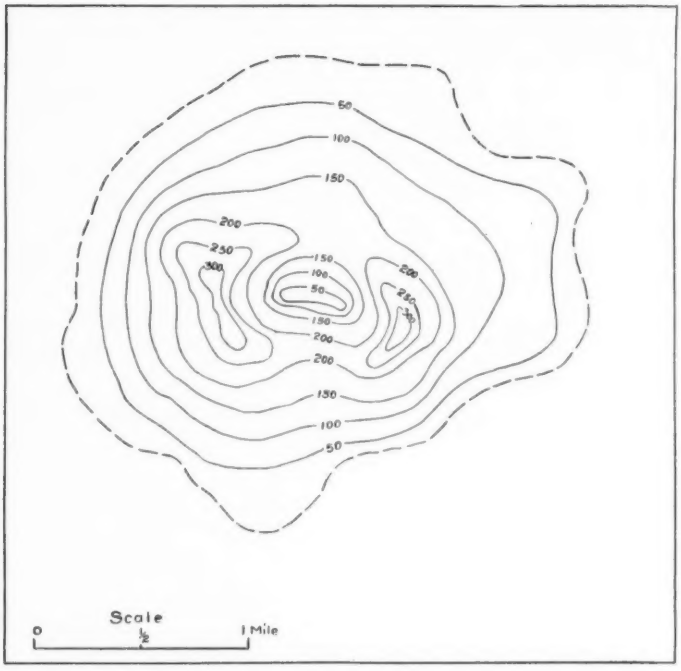

FIG. 6.—Isopach map showing the thickness of the Chattanooga shale. The average thickness of the shale outside the dashed line is 20 feet. Contour interval is 50 feet.

the result of the high-angle dip to the north of magnetic lines of force in the earth's surface.)

Small, approximately circular depressions are known to occur at many localities on the surface of the earth. Such depressions, similar in size, shape, and depth to the Flynn Creek disturbance, have been attributed to the following origins: (1) fall of a meteorite, with the resulting impact and explosion crater; (2) local collapse of a cavern

roof; (3) salt domes; (4) local expansion by hydration of anhydrite; (5) natural gas explosion; and (6) crypto-volcanic (gas and steam) explosion. The central uplift of 500 feet in the Flynn Creek area that raised Lowville limestone up to the level of the Leipers formation

FIG. 7.—Structure contour map on top the Chattanooga shale showing post-Chattanooga structure due to settling and compaction. Contour interval is 50 feet.

definitely eliminates a meteorite crater or the collapse of a cavern. The absence of any known salt or anhydrite deposits in this region makes an origin by salt-dome intrusion or by expansion of anhydrite unlikely. The stratigraphic horizon (Lower Ordovician) makes it improbable that sufficient natural gas occurred deep enough below the pre-Chattanooga surface to have blown out a crater. Further, although natural gas occurs at other localities in great quantities

and under enormous pressure, it has never been known to have formed such a crater by natural explosion. Crypto-volcanic explosion is the only possible origin among those listed that cannot be readily eliminated.

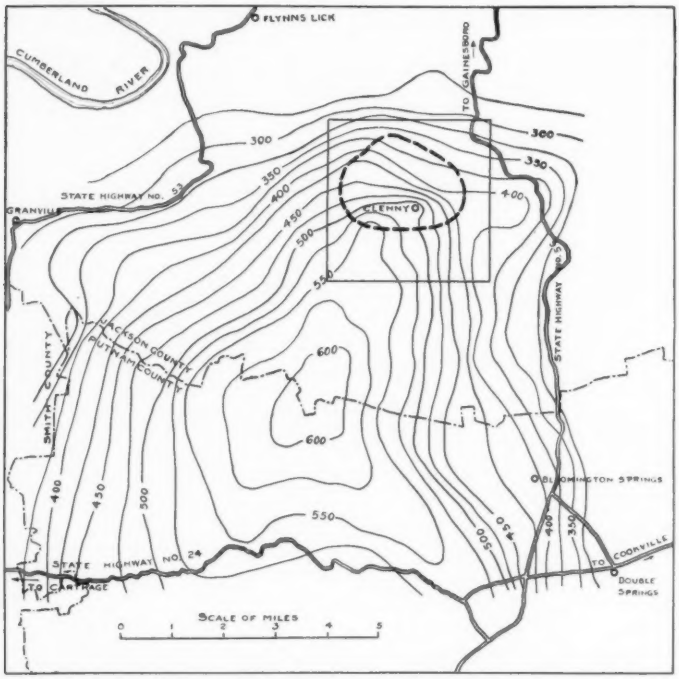

FIG. 8.—Isogammal map of parts of Jackson and Putnam counties, Tennessee. The lines are drawn through points of equal magnetic intensity. The Flynn Creek area (Fig. 2) is indicated by the square, the heavy broken line showing the position of the area of intense brecciation. Interval between lines is 25 gamma.

Lusk[9] believed the Flynn Creek disturbance to have been "formed by the collapse of the roof of an irregular branching cavern or series of caverns. The fragmentation induced by collapse, together with the slope wash of talus towards the lines of collapse would form the conglomerate-breccia."

[9] Lusk, "Geology and Oil and Gas Resources of the Gainsboro Quadrangle, Tennessee," *loc. cit.*, p. 35.

He discussed possible volcanic origin as follows:

> The question arose as to the possibility of post-Chattanooga diastrophism restricted to this small area. There might have been subsidence, or perhaps uplift followed by subsidence which deformed the shale so that at this one place it is exposed in the bed of Flynns Creek, whereas both upstream and downstream from this locality its base is high on the sides of the valley. The diastrophism might have proceeded in such a way as to produce the brecciation and high dips in the limestone. Bucher has described a circular area of intense folding and faulting in Adams County, Ohio, and refers it to the class of "crypto-volcanic" structures. However, the thickness of the shale, and the absence of folding or brecciation of the shale and overlying beds, which are structurally conformable, indicate that the Chattanooga shale and later formations could not have been affected by such diastrophism. Moreover, the actual exposure of the contact of the shale and breccia shows an irregular erosion surface which is not paralleled by the contact of the Chattanooga shale with the overlying green shale at the base of the Ft. Payne chert. The conglomeratic nature of the breccia, and the absence of veins or dikes of possible igneous origin discourage the view that sub-surface vulcanism may have been the cause of the diastrophism, either before or after Chattanooga time.

From this it is seen that Lusk correctly eliminated any post-Chattanooga volcanic origin, but he eliminated pre-Chattanooga volcanic origin only on the basis of (1) the "conglomeratic nature of the breccia" and (2) the absence of veins or dikes of igneous origin. The present writers did not find sufficient evidence of rounding, or "conglomeratic nature," of the breccia, and hence believe that all breccia but that designated as talus breccia resulted from the mechanical fragmentation of limestone and subsequent cementation. Also, they cannot accept the absence of veins or dikes of igneous origin at the surface as sufficient to eliminate a possible volcanic origin.

Bucher has described striking structural features in Adams County, Ohio;[10] Stewart County, Tennessee;[11] and Shelby County, Kentucky.[12] In his comparison of these structures with the Steinheim basin and Ries basin in Germany,[13] the origin of which had previously been attributed to volcanic activity, Bucher conclusively shows them to be homologous types of structure.

[10] *Op. cit.*, pp. 7–10. [11] *Ibid.*, pp. 12–16.

[12] *Ibid.*, pp. 5–7; "Geology of the Jeptha Knob," *Ky. Geol. Surv.*, Ser. 6, Vol. XXI (1925), pp. 193–237.

[13] W. Kranz, "Begleitworte zur geognostischen Spezialkarte von Wurttemberg," *Atlasblatt Heidenheim* (Stuttgart, 1924), Auflage 2, pp. 52–105.

The present writers believe that the Flynn Creek disturbance is similarly of volcanic origin and should be classed in the general group of crypto-volcanic structures. It is believed that (1) the small, circular central uplift of approximately 500 feet, (2) the intense brecciation of the limestone, (3) the intrusive character of the breccia, and (4) the shattering and jumbling of limestone blocks could have been caused only by a relatively rapid, deep-seated volcanic explosion accompanied by a gas explosion near the surface. The features are diagnostic of the examples of crypto-volcanic structures described by Bucher. It is very striking that the limestone in this area shows no evidence of hydrothermal alteration other than marked recrystallization.

The dip on the south flank in toward the center of uplift and the associated overturning and thrusting-away from the center are believed to be homologous with similar features in the Ries basin, Germany, where some of the formations that occupied the center were shoved, or thrust, radially outward from the center by the force of the explosion.[14] It is also thought that the closed, topographic depression on the pre-Chattanooga erosion surface was a crater formed by explosion. Similar explosion craters, or funnels, characterize the Ries basin and the nearby pit craters (*Randecker Maar*) associated with circular pipes and filled with basic tuff and fragments of the country rock.[15] In the Steinheim basin it is probable that a shallow saucer-shaped explosion funnel was first formed and that the central hill was formed a short time later by a second, weaker explosion.[16]

The Flynn Creek crater, which coincides in position with the central uplift, had two deep depressions below the general level of the floor (Fig. 5). The writers believe that the deep, pre-Chattanooga topographic basin was an actual crater formed by explosion, because (1) rocks were forced upward as much as 500 feet, giving an excess of material near the surface that must be accounted for, and (2) the only possible means of excavation by agents of erosion

[14] Bucher, "Cryptovolcanic Structures in the United States," *loc. cit.*, pp. 21–22.

[15] A. Bentz, "Der heutige Stand des Reisproblems," *Zeitschr. Deutsch. Geol. Ges.*, Vol. LXXXI (1929), pp. 70–75; W. Ahrens, "Die Tuffe des Nördlinger Rieses und ihre Bedeutung für das Gesamtproblem," *ibid.*, pp. 94–99.

[16] Bucher, "Cryptovolcanic Structures in the United States," *loc. cit.*, p. 3.

is by sinkhole solution, as the topographic basin was completely closed, having no outlet. There are no evidences of sinkhole solution in the pre-Chattanooga rocks of this region; and, also, it is believed that elevation above sea-level of central Tennessee during the Maysville-Chattanooga interval was never sufficiently high to permit the erosion of a 300 foot sinkhole, of which this would be the only known example.

At the time of explosion a great volume of limestone blocks was blown out of the crater, some falling back into the crater, and the majority being scattered around the margin of the crater within a radius of several miles.[17] Postexplosion and pre-Chattanooga erosion succeeded in removing all traces of the "cone" of brecciated limestone that surrounded the crater, as the writers were unable to find fragments of breccia at the base of the Chattanooga shale around the crater. The time interval between the explosion and the deposition of the Chattanooga shale must have been sufficiently long for the removal of the débris from the vicinity of the crater. On the other hand, the explosion could not have occurred long before Chattanooga time, as the crater would probably have been filled with sediments. Since the time necessary for the removal of the unconsolidated débris would not have been long geologically, and since the crater probably would have been filled during long existence as an open depression, the writers believe that the explosion shortly preceded the deposition of the Chattanooga shale, possibly in early Kinderhook or late Devonian time.

The abnormal thickness of the Chattanooga shale is attributed to the filling of the crater during the deposition of this formation. The syncline in the top of the black shale and in the overlying Fort Payne chert is believed to have resulted from three causes:

1. Initial dip upon the steeply sloping walls of the crater and the central hill would undoubtedly have been an appreciable factor in explaining the high dips (as high as 25°) present in the base of the thick Chattanooga shale, as contrasted with the much lower dips in the top of the shale.

[17] In the Ries basin, explosion breccia is found as far as 15.8 miles from the edge of the basin. Isolated blocks which are assumed to have been derived from the Ries basin have been found 37 miles from its edge. See W. H. Bucher, "Volcanic Explosions and Overthrusts," *Trans. Amer. Geophys. Union, 14th Ann. Meeting* (1933), pp. 238–42, esp. p. 240.

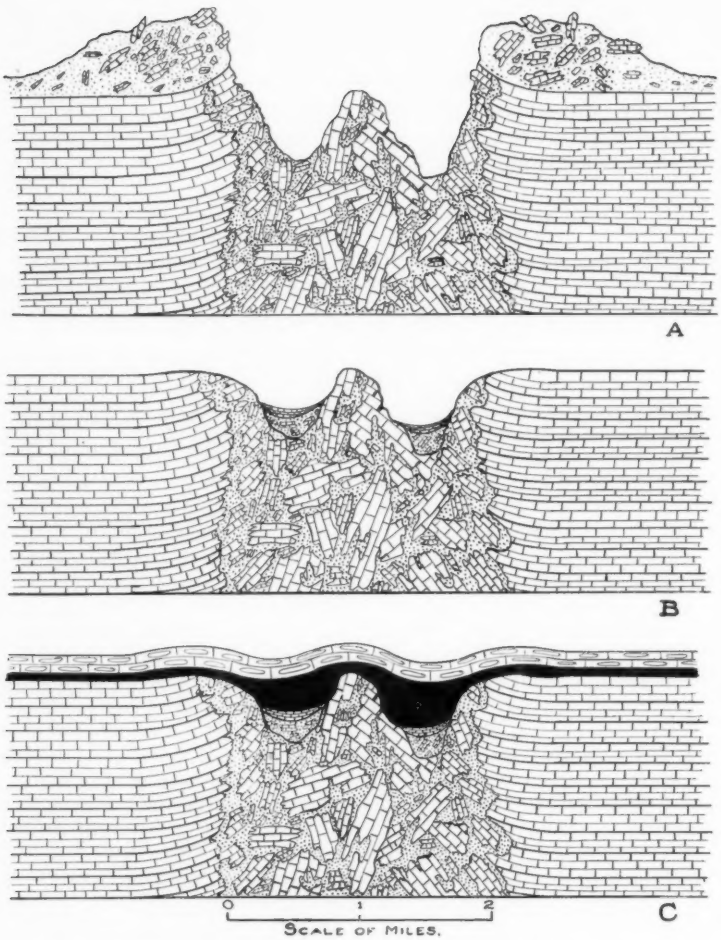

FIG. 9.—Diagrammatic restorations of a section across the Flynn Creek disturbance: *A*, shortly after the explosion (post-Leipers; pre-Chattanooga); *B*, after erosion had removed all fragmental débris from the vicinity of the crater, and after the deepest parts of the crater had been filled with talus débris and later fresh-water bedded breccia; *C*, sufficiently long after the deposition of the Chattanooga shale and the Fort Payne chert for subsurface readjustment of blocks and for compaction of the Chattanooga shale to have produced appreciable sagging in these two formations.

2. Compaction and proportional thinning of the 20 feet of shale around the crater and of the 250–300 feet within the crater would form an appreciable synclinal basin in the top of the shale.

3. Subsurface collapse and settling due to deep-seated readjustment would undoubtedly have resulted in post-explosion synclinal sagging and possibly faulting due to differential settling. No post-Chattanooga faulting of sufficient throw to map was found by the writers.

Summary of Events in the Flynn Creek Area

1. Deposition of the Lowville, Hermitage, Catheys, Cannon, and Leipers formations. If younger Ordovician or Silurian formations were deposited, they were removed by pre-explosion erosion, as the Leipers is the youngest formation involved in the explosion.

2. A volcanic explosion, blowing out a crater 300 feet deep and 2 miles in diameter, and piling up limestone débris in the vicinity of the crater at some time between the deposition of the Leipers and Chattanooga formations (Fig. 9*A*). This explosion preceded the deposition of the Chattanooga shale sufficiently to permit removal of all blocks of limestone around the crater (Fig. 9*B*).

3. Accumulation of talus in the deeper part of the crater, resulting from gravity rolling and slope wash of rock débris into the crater (Fig. 9*B*).

4. Formation of a fresh-water, ring-shaped lake that occupied the crater and surrounded the central hill leaving it an island. In this lake was deposited as much as 12 feet of bedded breccia (Fig. 9*B*).

5. Transgression of Chattanooga sea, which filled the crater with 250–300 feet of black mud and covered the surrounding region with 20 feet of similar sediments (Fig. 9*C*).

6. Deposition of Fort Payne chert.

7. Subsequent local synclinal sagging caused by compaction of the underlying Chattanooga shale and by subsurface readjustment following explosion.

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PERIGLACIAL LANDSLIDE TOPOGRAPHY OF CANJILON DIVIDE, RIO ARRIBA COUNTY, NEW MEXICO[1]

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ABSTRACT

On the higher parts of the Canjilon Divide, New Mexico, occur thick deposits of glacial till which are believed to represent the maximum extension of the oldest, or Cerro, glaciation of the San Juan Mountains of Colorado. On the lower parts of the divide and on the gently sloping upland west of it, there are numerous discontinuous and comparatively thin bodies of till which form a distinctively hummocky topography. The topographic forms and the distribution of till are satisfactorily explained only as having been formed by landsliding of bodies of till derived from a thick till mantle capping the remnants of the high-level erosion surface. The sliding took place on gentle slopes, now stable, and is believed to have resulted from intense frost action during glacial stages of the Pleistocene, when periglacial climatic conditions must have prevailed. Two widely separated periods of sliding are suggested by the comparative sizes and the distribution of the various landslide masses. These are correlated, tentatively, with the Durango and the Wisconsin glacial stages of the San Juan Mountains.

INTRODUCTION

Canjilon Divide is located in Rio Arriba County, north-central New Mexico. It lies west of the Rio Grande depression, in the extension of the San Juan upland of Colorado south into New Mexico. The divide, as considered in the following pages, lies within the Abiquiu quadrangle of the United States Geological Survey, and extends about 7 miles south from the northern boundary of the quadrangle, which is some 35 miles south of the Colorado–New Mexico state line (Fig. 1).

The features of Canjilon Divide are of more than local interest. The glacial till which covers the higher parts of the divide represents the southernmost extension of an ancient piedmont glacier of the San Juan Mountains, and thus contributes to a more complete picture of the regional glacial history. The associated periglacial features attest the effects of glacial climates beyond the areas actually glaciated during later stages of the Pleistocene, and add to the small

[1] One section of a doctorate thesis, entitled "The Tertiary and Quaternary Geology of the Abiquiu Quadrangle, New Mexico," submitted at Harvard University, 1935.

but growing number of examples of periglacial phenomena reported from North America.[2] The presence of the till is significant also in providing a starting-point in estimating the age of erosion surfaces in the Rio Grande Valley.

The present study was made during the summer of 1934, in the course of a more general survey of the geology of the Abiquiu quadrangle. The base map employed was the advance sheet of the Abiquiu quadrangle, which, in this area, was compiled by N. H. Darton from forest reconnaissance topographic maps of the Carson National Forest, surveyed by pacing and aneroid barometer. This map, on a scale of 1:125,000 and with a contour-interval of 100 feet, shows the general form of the topography satisfactorily but is rather inaccurate as to elevations. Figure 1 is reproduced from this base map.

GENERAL DESCRIPTION OF THE DIVIDE

Canjilon Divide is a north-south ridge ranging from 9,000 to 10,000 feet in elevation. On the east it descends steeply some 1,500 feet to El Rito Creek, a vigorous tributary of the Chama River, which itself is a tributary of the Rio Grande. West of the crest of the divide, a rolling upland slopes gently toward the Arroyo Canjilon, a less agressive tributary of the Chama, which meets that stream about 15 miles upstream from its junction with El Rito Creek. This gently sloping upland comprises the remnants of two distinct erosion surfaces. The higher, or No. 1, surface is now represented by three isolated remnants (*A*, *B*, and *C*, Figs. 1 and 2), which lie at elevations of approximately 9,500 to 10,000 feet. These remnants are capped by glacial till of undetermined maximum thickness.

Extending fanwise outward from the high-level remnants, there are broad remnants of a lower, or No. 2, surface, which slopes gently to the west and which, except at the marginal portions, is only moderately dissected. At locality *B* (Figs. 1 and 2) the two surfaces are separated by a comparatively steep east-west scarp some 500 feet high (Fig. 3). Elsewhere there are moderately steep transition slopes leading from the lower surface up to the higher. The existence of extensive remnants of both surfaces is due entirely to the fact that

[2] Cf. Kirk Bryan, "Glacial Climate in Non-glaciated Regions," *Amer. Jour. Sci.*, Vol. XVI (1928), pp. 162–64.

stream erosion in this upland region has been very sluggish in its response to changes in baselevel and has lagged many cycles behind the Chama Valley.

Incised below the level of the No. 2 surface, there are shallow youthful valleys of stage No. 3. These valleys are at present cutting

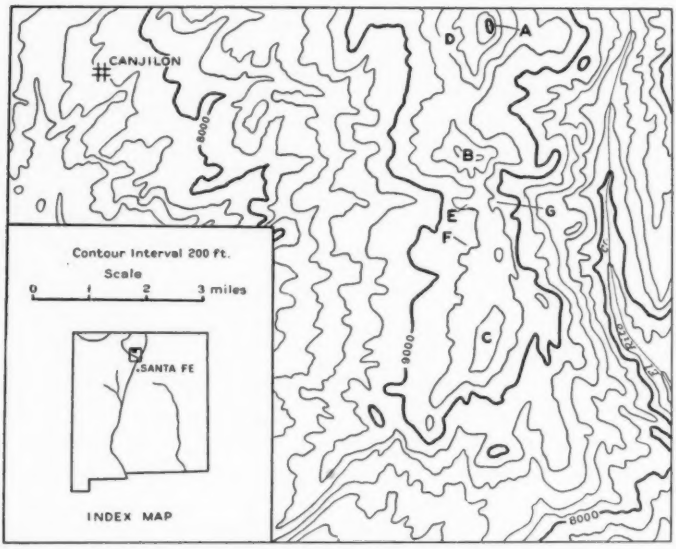

FIG. 1.—Generalized topographic map of the Canjilon Divide area. The index map shows the location of this area in the Abiquiu quadrangle, and also the location of the Abiquiu quadrangle in the state of New Mexico. The letters refer to localities discussed in the text: *A*, *B*, and *C* are the localities at which glacial till occurs in place; *D*, *E*, and *F*, the localities at which periglacial landslide topography is particularly well displayed; and *G* is a locality of modern landsliding. At locality *E* a detailed topographic map was made (Fig. 5).

headward into surface No. 2. To the west they widen out and form parts of a prominent erosion surface.

The surfaces described above are all cut on the Mancos shale, of upper Cretaceous age, which dips gently to the west.[3] This shale weathers readily to a colloidal clay, very slippery when wet.

[3] N. H. Darton, "'Red Beds' and Associated Formations in New Mexico," *U.S. Geol. Surv. Bull. 794* (1928), Fig. 68 and Pl. 37.

OCCURRENCE OF THE TILL

Previous observations.—In 1901 Stone[4] wrote that "the main southeast spur of the San Juan Mountains passes into New Mexico as the Conejos range. This range falls rapidly toward the south and was glaciated for not more than 30 to 50 miles south of the Colorado–New Mexico line." Stone did not state the age of the glaciation, however, nor did he cite any specific localities. Antevs,[5] in mapping the Pleistocene glaciation of Western North America, followed Stone in showing this glaciation of north-central New Mexico, and was also noncommittal as to its age.

Occurrence and description of the till.—In the course of the present study, the glaciation mentioned by Stone was found to be represented on Canjilon Divide by till deposits of considerable thickness, which must once have been of wide extent. At two localities, *A* and *B* on the map (Fig. 1), particularly good exposures were found. At these localities the till was seen to be composed of angular to subrounded boulders and rock fragments of all sizes up to at least 2 feet in length, mingled indiscriminately in a very compact gritty matrix. A few boulders approaching the "soled" shape characteristic of glacial abrasion were found, and also a few bearing marks suggestive of striations. The boulders were all derived from pre-Cambrian crystalline rocks, mainly quartzite, which do not appear in the immediate vicinity. At the northern locality (*A*, Fig. 1) they must have been transported a minimum distance of 8 miles, the distance north

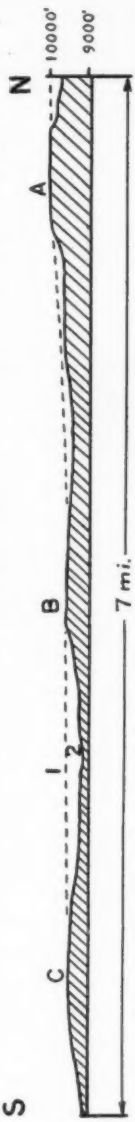

FIG. 2.—Topographic profile of Canjilon Divide (somewhat foreshortened), traced from a photographic panorama. Erosion surfaces are indicated by numbers and specific localities by letters, as in Figure 1.

[4] G. H. Stone, "Note on the Extinct Glaciers of Arizona and New Mexico," *Abst. Science*, Vol. XIV (1901), p. 798.

[5] Ernst Antevs, "Maps of the Pleistocene Glaciations," *Geol. Soc. Amer. Bull.* 40 (1929), Fig. 7.

to the nearest outcrop area of pre-Cambrian rock, as shown on the state geological map.[6]

The condition of the till is such as to suggest long-continued weathering. In the upper 15 feet the till is less compact than in the main mass, and the matrix ranges from sandy to clayey. Numerous boulders throughout the entire section are thoroughly rotten, par-

FIG. 3.—View of east-west scarp at locality *B*, looking north across a part of locality *E*. The garland-shaped bouldery area in the foreground is a secondary landslide mass of till, derived from a primary landslide mass represented as the larger bouldery area upslope. The heap of débris on the garland-shaped mass marks the location of the test trench. Near the lone pine tree on the horizon is a 35-foot exposure of till, freshened by modern landsliding.

ticularly those of schist and gneiss, which almost invariably crumble when tapped with the hammer. Many boulders of quartzite, most resistant rock in the region, fall apart almost as readily as those of schist.

The presence of the till at locality *C* and between the few good exposures is indicated by a characteristic float of quartzite cobbles and

[6] N. H. Darton, "Geologic Map of New Mexico" (Washington: U.S. Geol. Surv., 1928).

boulders, including some erratics up to 8 feet in length. The minimum thickness of the till is 130 feet at locality *A* and 35 feet at locality *B*, the base being concealed at both places.

Age and correlation.—The great age of the till on Canjilon Divide is indicated both by the state of weathering, already referred to, and by the topographic relations, considered in the following paragraphs.

1. Deep dissection through successive cycles: The existing till caps remnants of the highest erosion surface of the region, probably equivalent to the San Juan peneplain of southern Colorado.[7] To the east and adjacent to these high-level remnants lies the valley of El Rito Creek, more than 1,500 feet deep, with indications of having passed through three or more cycles of erosion since the initial dissection of the till-covered surface. This erosional history indicates a long interval of time since the deposition of the till, extending back to the earlier stages of the Pleistocene.

2. Topography on which till was deposited: The topographic position of the till indicates that it was laid down on a widespread surface of low relief, and that it probably represents the deposit of a piedmont rather than of a valley glacier. Furthermore, the thickness of the till and the known minimum distance of transportation testify to the vigor of glacial action and point to a stage of maximum glacial extension. These conditions are comparable only with those of the oldest, or Cerro, glacial stage in the San Juan Mountains to the north, when piedmont glaciers extended many miles beyond the limits reached by the valley glaciers of later stages.[8]

The weight of evidence thus indicates that the till on Canjilon Divide can be no younger than the Cerro till of the San Juan Mountains in Colorado, and that it marks the southernmost extension of a widespread piedmont glacier which once covered the San Juan region.

DETAILED FEATURES OF THE TOPOGRAPHY

The remnants of surface No. 2 on the western slopes of Canjilon Divide are characterized by a conspicuously hummocky topography,

[7] W. W. Atwood and K. F. Mather, "Physiography and Quaternary Geology of the San Juan Mountains, Colorado," *U.S. Geol. Surv. Prof. Paper 166* (1932), chaps. iv, vi, and vii.

[8] *Ibid.*, pp. 82–83.

quite in contrast with the smoothness of the till cover on remnants of the higher surface. The presence of some till on the lower surface is indicated by actual exposures at the very edge of the divide, where its thickness is less than 10 feet, and by numerous detached bouldery areas at other places. At localities *D*, *E*, and *F*, of Figure 1, the hummocky topography is especially well developed. At locality *D*, discontinuous, boulder-covered mounds and ridges, up to 15 feet or more in height, and trending roughly north and south, present an appearance suggesting, in miniature, the topography of a terminal moraine (Fig. 4). The area intervening between the hummocky belt

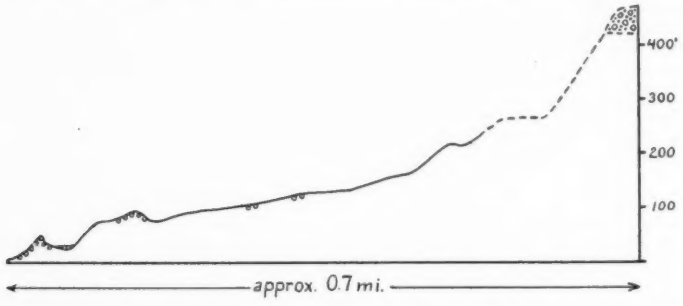

FIG. 4.—Profile across locality *D* to locality *A* on Canjilon Divide, showing the pseudo-morainic topography. Bouldery areas are indicated by groups of circles. Profile is based on pace and hand-level traverse by H. H. Mellus.

and the higher till-capped butte slopes rather gently and more or less regularly toward the west, has no integrated drainage, and is characterized by a marshy type of vegetation. At the outer edge of the hummocky belt there are several small ponds. Beyond these there are steeper slopes, with youthful valleys working headward into the hummocky belt.

Locality *F* differs from *D* in that the bouldery mounds and associated features are spread out all along the slope rather than being concentrated in a definite belt. The same is true, though to a lesser extent, at locality *E*, which was selected for more intensive study. At this locality a plane-table topographic map was made (Fig. 5), on a scale of 100 feet to the inch and with a contour interval of 2.5 feet. The panoramic view, Figure 6, shows the mapped area in its natural setting. Inspection of the map and panorama shows that small ponds

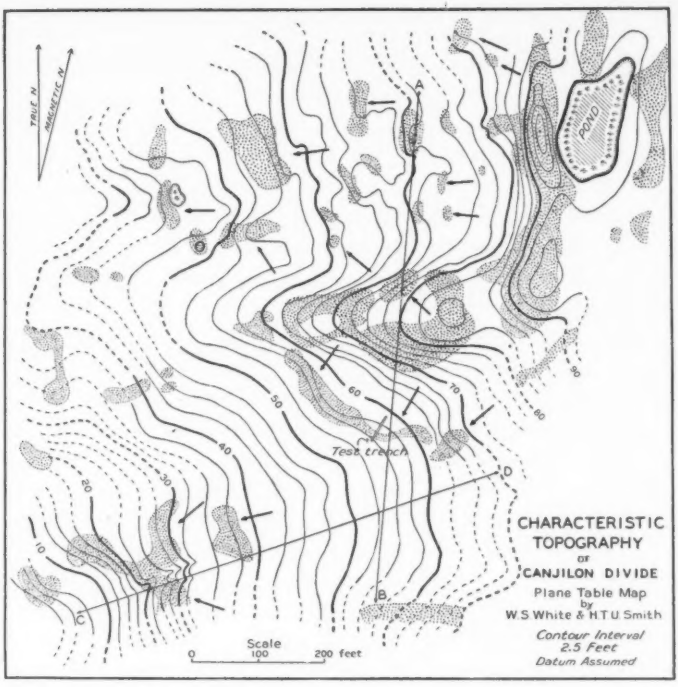

FIG. 5.—Detailed topographic map of locality *E* on Canjilon Divide. The stippled areas represent bouldery outcrops which are interpreted to be partly buried landslide masses of till. The arrows show the direction of sliding postulated for particular masses.

FIG. 6.—Panoramic view looking south across locality *E* on Canjilon Divide. Localities *C* and *F* are in the background. The map area of Figure 5 lies in the right-hand half of the panorama.

are present, that drainage is unintegrated, and that bouldery hummocks are a prominent feature of the landscape. These hummocks, on the basis of form and size, are divisible into two groups. The first group, comprising the larger and more irregularly shaped hummocks, is restricted, for the most part, to the higher ground—to remnants of surface No. 2. The second group comprises the smaller lunar, garland-shaped, and elongate boulder-banked and boulder-covered forms, with undrained depressions on the upslope side in some places. The hummocks of this group occur mainly on the lower ground—in shallow valleys slightly below the true level of surface No. 2. These valleys show no distinct or definite drainage channels and have rather the form of a series of miniature open basins strung out along the slope, descending steplike from one to another. The average gradient of these irregular slopes is about 4° or 5°.

The hummocky topography is unrelated to present-day processes of erosion, which include sheet wash, stream incision, and landsliding. Sheet wash was observed to take place within the area during heavy rainfall. It serves to transport a small amount of fine débris, and thus to fill up the remaining undrained depressions. Stream incision is working headward into the western margin of the hummocky area but has, thus far, made little headway at locality *E*. Modern landsliding is taking place on a small scale and is confined to the steeper slopes (of the order of 16°) east of the divide and on the southern side of the east-west escarpment (Fig. 3). Its effects are displayed in small ponds (locality *G*), in fresh scars still free of vegetation, and in cirque-like re-entrants which notch the rim of the escarpment (locality *B*).

LANDSLIDE ORIGIN OF THE HUMMOCKY TOPOGRAPHY

Casual observation of the hummocky topography suggests three possible ways in which it might have originated: (1) by direct deposition of the till, perhaps with some subsequent dissection; (2) by reconcentration of boulders derived from weathering or erosion of the till, as by the processes involved in the formation of altiplanation terraces; or (3) by bodily movement of large masses of till. As shown in the following pages, field study demonstrated that the first two of these possibilities are untenable, and pointed instead to landsliding

of large masses of till as the only plausible explanation for the existing topography.

The smaller bouldery hummocks.—The evidence for landslide origin of the group of smaller bouldery hummocks is more obvious than that for the group of larger hummocks, and therefore is discussed first.

1. Topographic position incompatible with hypothesis of morainic origin: It is well known that morainic topography, in general, is preserved only in those areas glaciated during the later stages of the Pleistocene, subsequent to which there has been but slight dissection. Any glaciers so recent as to have left well-preserved morainic forms in the area discussed could not possibly have covered even the lower parts of the divide, but would have followed the adjacent valleys, which are far too large to have been excavated since late Pleistocene time. Neither in these valleys nor elsewhere in the area are there any records of recent glaciation.

2. Topographic form different from that of altiplanation terraces: The hummocky features here discussed bear a general resemblance to the altiplanation terraces described by Eakin[9] and others.[10] Both types are characterized by an elongate form, a boulder-banked face, and a more or less steplike arrangement. The features on Canjilon Divide differ from true altiplanation terraces, however, in two respects: (*a*) the stone-banked face has a much gentler slope than that reported by Eakin as characteristic of altiplanation terraces, which is near the angle of repose of the cobbles or boulders; and (*b*) the true altiplanation terraces are not accompanied by depression forms on the upslope side.

3. Compact till revealed at shallow depth by test trench: A test trench 45 feet long and from 1.5 to 4.5 feet deep was dug across the garland-shaped mass pictured in Figure 3 (see also map, Fig. 5). The section exposed in this trench is shown in Figure 7*A*. The boulders which appear at the surface were found to lie in a rather

[9] H. M. Eakin, "The Yukon-Koyukuk Region, Alaska," *U.S. Geol. Surv. Bull. 631* (1916), pp. 76–82.

[10] Ernst Antevs, *Alpine Zone of Mt. Washington Range* (Privately printed, Auburn, Me., 1932), chap. iv; R. J. Russell, "Alpine Land Forms of Western United States," *Geol. Soc. Amer. Bull. 44* (1933), pp. 927–50.

loose soil matrix, which, at a depth of about 1 foot, grades down into compact, massive till, with boulders so tightly imbedded in the matrix as to make further excavation slow and arduous. The top of the till apparently dips into the slope of the topography. Overlapping the till on the upslope side is a fill of loose-textured, black soil, which grades down into more compact, lighter-colored clay. This

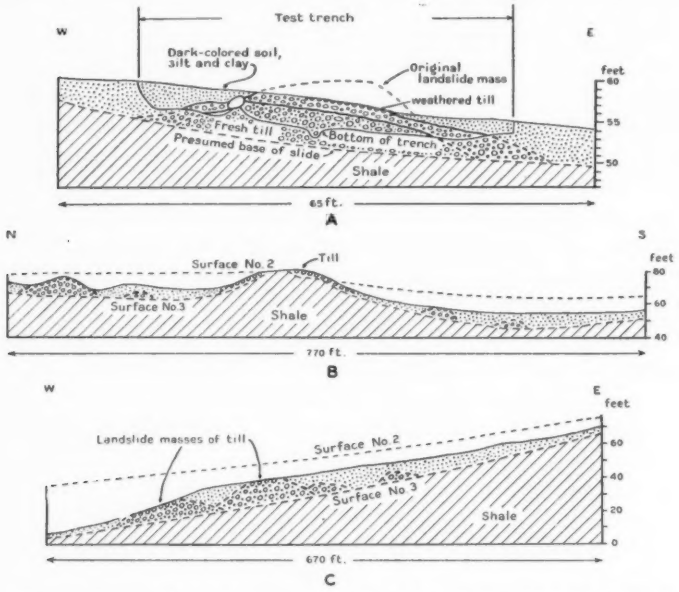

FIG. 7.—Cross sections. *A*, cross section through the test trench; *B*, profile and inferred subsurface relations along line *AB*, Figure 5; *C*, profile and inferred subsurface along line *CD*, Figure 5.

fill was evidently deposited in the enclosed depression which originally lay behind the mass of till. Filling may have been effected by solifluction in the earlier stages, certainly by rainwash in the later stages. Overlapping the till on the downslope side is a similar fill, presumably deposited in the depression behind other landslide masses which blocked the valley farther down (Fig. 7*C*). The similarity in form between the other bouldery hummocks and the one in which the trench was dug is considered sufficient reason for believing

that they also represent masses of till, although no additional trenches were dug.

4. Orientation of the smaller masses suggests derivation from the larger: The orientation of many of the smaller detached masses of till is such as to suggest that they moved to their present positions from the larger masses. The arrows in Figure 5 indicate directions of movement postulated for particular masses.

5. Similarity to modern landslides: The form of the till masses, and the association with depression forms on the upslope side in some places, is similar to that of the modern slides which have been described. The latter differ, however, in being of smaller size, in having moved on much steeper slopes, and in being marked by fresh scars, many still free of vegetation. The difference with respect to slope is such as to indicate a difference in process. At present the main process of sliding is probably direct gravitational settling on slopes oversteepened by erosion, and is facilitated by saturation of weathered shale underlying the till, with frost action as a possible contributing factor. It is obvious that this process is not, and could not have been, effective on the much gentler slopes over which the older slides moved. It is probable, rather, that sliding in the former period was caused by intense frost action accompanying a formerly more severe climate. In a later section of this paper the probable mechanism of this type of sliding is discussed in detail.

6. Prevalence of conditions favorable for sliding: The till is a massive and relatively competent formation, whereas the shale on which it rests is relatively incompetent and dips with the slope of the topography. The shale is readily weathered and, in the weathered condition, when wet, forms a plastic, yielding mass.

From the foregoing considerations, it may be concluded that landslide origin of the group of smaller hummocks is indicated by both negative and positive evidence—negative in excluding alternative possibilities, and positive in revealing the form, structure, and general relations indicative of landslide bodies.

The larger bouldery hummocks.—The origin of the larger till masses from which the smaller ones were derived remains to be considered. In the following paragraphs it is proposed to show that these could not have been morainal deposits and that landsliding from a former

till-capped remnant of surface No. 1, since removed by erosion, remains as the only valid explanation for their presence.

In the first place, it is clear that the till on the crest of the divide at locality *E*, on surface No. 2, could not have been deposited by the same ice sheet which left the till on surface No. 1 at locality *B*, immediately to the north. The hummocky topography of the lower till stands in contrast to the smooth surface of the higher till, and is not readily explainable if it is assumed that the till at both places was deposited by the same glacier at the same time. The 500-foot scarp in soft shale, separating the till of localities *B* and *E* (Fig. 3), presents additional difficulties to the hypothesis of contemporaneous deposition. The distribution of till at the top and bottom of the scarp, and on one of the re-entrants halfway up the slope, is such as to demand that the scarp must have been left with essentially its present steepness when the ice retreated, *if* the till below is contemporaneous with the till above. This is manifestly improbable, for glacial erosion on uniformly soft rock tends invariably to produce gentle slopes, and the thickness of till here indicates that glacial action must have been vigorous. Origin of the scarp by faulting of an original till-covered surface is equally improbable, for no faults of the necessary recency, or having the required trend, are known in this district.

In the second place, it is unlikely that the till on the lower surface could have been deposited by a later glacier. The possibility of movement directly from the north is excluded by the undisturbed character of the higher till, by the steepness of the intervening scarp, as noted above, and by the fact that the upland to the north was undoubtedly a divide between two valleys. Movement down the valley to the east is also improbable, for this valley is not through-going to the glaciated area of the San Juans, and there is no reason for believing that it ever was. Even if it be imagined that this valley was glaciated during the period in question, it is very doubtful whether the known volume of morainic débris could have been carried to its present position by overflow across the divide. Finally, it is no less difficult to picture deposition by a glacier advancing down the broad, through-going valley to the west. Such a glacier could have reached locality *E* only by turning a corner around the upland remnant of surface No. 1 and by moving uphill for a considerable

distance in a direction highly oblique to the trend of the main valley. This is obviously improbable.

Landsliding alone remains to account for the form and distribution of the till on surface No. 2. The slope of the topography and the orientation of the till masses indicate that the sliding must have been from the east. However, there is no visible source of till in that direction; hence it must be inferred that the till was derived from an upland which formerly extended east of the present divide and which since has been removed by erosion as the divide was pushed westward. This upland area was undoubtedly part of an unbroken remnant of surface No. 1, continuous from locality *A* to locality *C*.

The discussion thus far, although based primarily on locality *E*, applies also to localities *D* and *F*. At locality *D* the effects of the ancient sliding are even better preserved and the parent-upland remnant has not yet been eroded away. Here the total distance of sliding must have been more than half a mile.

AGE OF THE LANDSLIDING

Evidence of present stability.—A careful study of the landslide topography fails to disclose any indications, however slight, of present-day movement. The vegetative cover is continuous and undisturbed, with trees growing on and downslope from the landslide masses, in such positions as to be easily disturbed by any continuation of the movement. The depressions behind all of the smaller landslide masses are either completely filled or nearly filled; and, if the rate of filling is to be estimated from the effects of sheet wash observed during several heavy downpours, it must be a very slow process. Finally, the valleys which are now working headward into the marginal portions of the landslide area show no indications of having been interrupted by any recent sliding.

Evidence of relative recency of the movement in part.—Although no movement is now in progress, the very nearly perfect preservation of the topographic forms, particularly of those associated with the group of smaller slides, does indicate that these forms were produced in the not very distant past. The larger ponds are not yet completely filled; and, save for the very marginal portions, the landslide area is entirely undissected. The state of preservation of the various fea-

tures is fully comparable with that of features produced during the Wisconsin stage of glaciation.

Reasons for postulating landsliding during a glacial period.—The present stability within the landslide area is evidently due to the cessation of some particular process, and not to the attainment of a state of equilibrium. Inspection of the map (Fig. 5) shows several places at which till masses, both connected and detached, lie on the edge of or part way down slopes considerably steeper than those on which a great part of the original sliding must have taken place. The fact that such masses have remained in place, whereas others, which started sooner, traveled distances of tens and even hundreds of feet, can be explained only as the result of the action of some process no longer in operation. Intense frost action is known to be an effective agent of transportation, and is, indeed, the only additional process which is available and which is adequate to have produced extensive movement on gentle slopes. A consideration of the effects of modern frost action, as observed in arctic and alpine regions, supports this conclusion. Högbom,[11] in Northern Sweden, shows photographs of boulders "plowing" down gentle slopes, under frost action, so vigorously as to push up a low swell in front, while leaving an open furrow to the rear. In Swedish Lapland, Högbom[12] noted the movement of great numbers of blocks 50–100 kilograms in weight over a bare rock surface sloping 5°–10°. From the summit of Mount Washington, in New Hampshire, Antevs[13] describes a block 25×20×6 feet which, under post-Wisconsin frost action, has moved a short distance down a 5°–10° slope. Numerous other, though less spectacular, examples of the efficacy of frost action as an agent of transportation might be cited from the writings of Högbom and other students of this type of phenomena.

During Wisconsin time it is highly probable that the climate in the vicinity of Canjilon Divide was of the periglacial type. A widespread ice sheet covered much of the San Juan region and reached southward at least as far as the Colorado–New Mexico state line,[14] coming

[11] Bertil Högbom, "Über die Geologische Bedeutung des Frostes," *Bull. Geol. Inst. Univ. of Upsala*, Vol. XII (1914), p. 350.

[12] *Ibid.*, p. 344. [13] *Alpine Zone of Mount Washington*, p. 48.

[14] Atwood and Mather, *op. cit.*, Pl. 3.

within 35 miles of the Canjilon Divide area. The conditions responsible for this glacier must certainly have had a marked effect on surrounding areas, producing a more severe climatic régime, akin to that of modern arctic and alpine environments. Under such conditions frost action must have been greatly intensified, providing a ready and powerful agent of transportation. From these considerations it is concluded that the smaller landslide masses date back to the periglacial conditions of Wisconsin time.

Reasons for postulating that part of the sliding took place in a more remote period.—In preceding pages it was shown that the presence of till on surface No. 2 at locality *E* is best explained as a result of landsliding from a once contiguous upland remnant of the No. 1 surface, since removed by erosion. Reconstruction of this former upland leads to an estimate of the amount of erosion which has taken place since the earlier period of sliding. Interpolation between localities *B* and *C* suggests that surface No. 1, at locality *E*, was probably 400–500 feet higher than the present crest of the divide. Assuming that the gradient of the transition slope between the two surfaces here was of the same order as it now is at localities *A* and *C*, extrapolation indicates that, at the time of the first sliding, the divide lay at least a quarter of a mile, perhaps even more than half a mile, east of its present position. This means that a great volume of material has been eroded away since sliding began, and that a great amount of valley-deepening and valley-widening has taken place. Such a degree of erosion is far greater than that known to have been effected during post-Wisconsin time generally, and, in particular, far exceeds the moderate amount of valley-deepening which marked post-Wisconsin time in the San Juan Mountains of southern Colorado, which offer the nearest basis for comparison. Thus it is concluded that the earlier period of sliding took place long before the advent of the Wisconsin glacial stage. For reasons similar to those presented above in connection with the later period of sliding, it is believed also that this earlier period of sliding probably accompanied a glacial stage. Correlation of this ancient period of sliding with the Durango glaciation of the San Juan Mountains is immediately suggested by the glacial chronology of the San Juan region, the Durango stage being next older than the Wisconsin. If this correlation is correct, the larger

scale and greater distance of sliding during the earlier period may be explained by the fact that the Durango ice was more extensive and lay nearer than the Wisconsin, and thus was probably represented on Canjilon Divide by periglacial conditions more severe or longer continued.

The hypothesis of two-cycle sliding, presented above, is admittedly based largely on inferential reasoning. It is believed, however, to provide the only adequate explanation for all of the field facts. It makes intelligible the apparent topographic unconformity between the surfaces on which the large-scale and the small-scale sliding took place. This may be interpreted as a result of the dissection of surface No. 2, together with some of its landslide masses, by normal stream erosion during the Durango-Wisconsin interglacial stage. The origin of the larger ponds is accounted for also, as described below.

Origin of the larger ponds.—The two ponds shown in Fig. 6 are rather large, and too closely related to the larger till masses to have been formed by the small-scale sliding, but are too well preserved to date back wholly to the earlier period of sliding. They may be satisfactorily explained, however, as having resulted from the combined effects of the two periods of sliding. The initial depressions were probably formed by sliding during the earlier period, and later were filled by sheet wash during the ensuing interglacial stage. This fill is now represented by the material exposed in the steeper southern bank of the pond shown in Figure 8. During the latter part of the interglacial stage, the present basins were outlined by moderate dissection of the previously formed fill. Later, in Wisconsin time, the outlets of the dissecting valleys were blocked by small-scale sliding, thus creating the existing depressions. In post-Wisconsin time these depressions were partly filled by sheet wash.

Provisional outline of physiographic history.—The sequence of events which played a part in modeling the present topography is shown diagrammatically in Figure 9, and listed as follows:

1. Formation of a widespread peneplain by subaërial erosion during late Tertiary time, continuing perhaps into the early Pleistocene.

2. Deposition of the Cerro (?) till, in considerable thickness, by the southernmost lobe of a piedmont glacier from the San Juan Mountains of Colorado.

3. Long-continued erosion to a lower baselevel, developing a second widespread, gently rolling surface, about 400–500 feet lower than the earlier peneplain.

4. Advent of the periglacial climate of the Durango (?) glacial stage, with intense frost action effecting large-scale sliding of till masses derived from unreduced remnants of the earlier peneplain.

FIG. 8.—Close-up view of the pond shown near the center of Figure 6. The bouldery mound on the right-hand side represents the earlier period of sliding. The low, steep bank toward the left-hand side of the picture represents the product of the first period of fill. Photograph by R. C. Goodwin.

5. A period of fill behind the landslide masses, followed by an interval of erosion to a lower baselevel. West of the divide the erosion of this interval was of very moderate extent, but east of the divide it was far more vigorous, in places removing all vestiges of the higher, till-covered surface, and shifting the divide itself westward.

6. Advent of the Wisconsin glacial stage, with return of periglacial conditions and renewal of frost action, leading to small-scale sliding of minor till masses derived from the earlier landslide masses. The heads of the dissecting valleys were thus blocked.

7. Post-Wisconsin sheet wash, fill, and stream incision, with continued vigorous erosion in the valley of El Rito Creek, east of the divide.

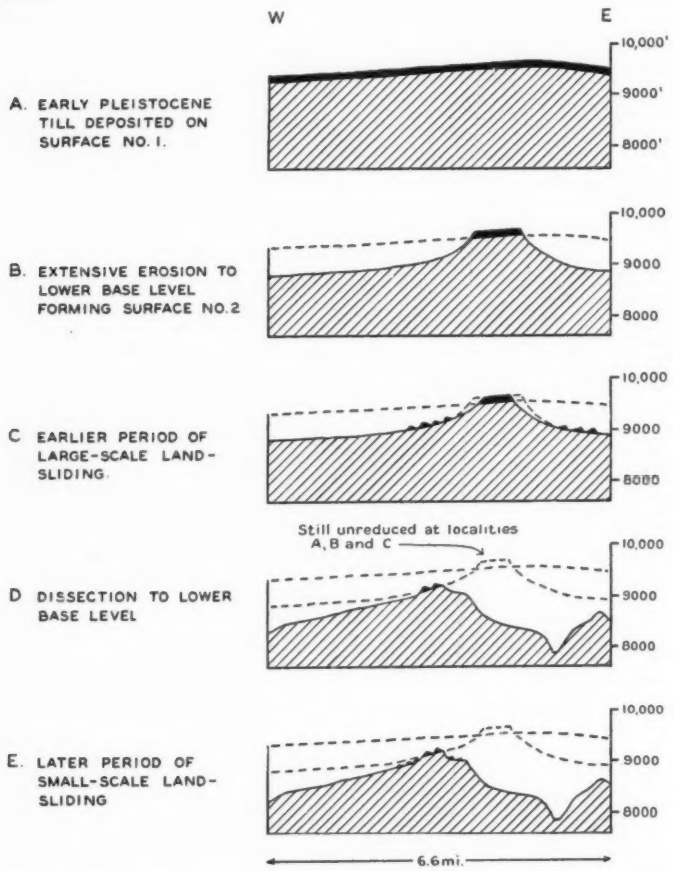

FIG. 9.—Diagrams illustrating principal events in the physiographic history of Canjilon Divide at locality *E*. Vertical exaggeration, 5×.

MECHANICS OF THE LANDSLIDING

The conclusion that landsliding on Canjilon Divide could have been due only to vigorous frost action, under periglacial conditions,

leads naturally to speculation concerning the mechanics of the process. Two distinct stages in the process are to be accounted for: (1) the detachment of numerous large bodies of till from the parent till sheet, and (2) the movement of these till bodies for long distances on gentle slopes.

Detachment of the till masses.—The Pleistocene landslide masses, particularly those of the first generation, were much larger and more numerous than those of the present. They were formed on slopes which, under present conditions, are essentially stable. These facts suggest that they were initiated through the agency of special processes which are no longer in operation—processes inaugurated by periglacial conditions.

One of the common conditions for landsliding, in general, is undermining or sapping by any one of several erosive processes. This applies especially to places where more resistant formations are underlain by less resistant formations. On Canjilon Divide the till constitutes the former; the underlying shale, the latter. At present, the process of undermining may be observed where steep slopes are being cut back into the till, but is absent on the gentler slopes which, at most places, lead down from remnants of the till mantle. During the glacial stages of the Pleistocene, however, conditions were undoubtedly different. Nivation and solifluction must have been important agents of erosion, as they are today under comparable climatic conditions in Greenland,[15] in Alaska,[16] on the higher peaks of our western mountains,[17] and elsewhere.[18] By these processes the shale, even on comparatively gentle slopes, was subject to attack and removal. Disrupted by frost weathering and saturated with water from gradually melting snow, it moved slowly downslope, perhaps in part by flowage in a semifluid state, and in part, perhaps, by reason of the

[15] W. E. Ekblaw, "The Importance of Nivation as an Erosive Factor and of Soil Flow as a Transporting Agency in Northern Greenland," *Proc. Nat. Acad. Sci.*, Vol. IV (1918), pp. 288–93; "The Importance of Solifluction," *Abst. Annals Assoc. Amer. Geogrs.*, Vol. XXI (1931), p. 121.

[16] S. R. Capps, "The Kantishna Region, Alaska," *U.S. Geol. Surv. Bull. 687* (1919), pp. 66–70.

[17] F. E. Matthes, "Glacial Sculpture of the Bighorn Mountains, Wyoming," *U.S. Geol. Surv. 21st Ann. Rept.* (1900), Part II, pp. 180–84; Russell, *op. cit.*

[18] J. G. Anderson, "Solifluction, a Component of Subaërial Denudation," *Jour. Geol.*, Vol. XIV (1906), pp. 91–112.

downhill component of repeated expansion and contraction attending freezing and thawing, as described in the following section of this paper.

It is probable that the till, meanwhile, yielded less readily to frost disruption. Standing on higher ground and in steeper slopes, it was much sooner freed from the effects of meltwater from lingering snowdrifts. In addition, its permeability, porosity, and particle size may have rendered it less susceptible to frost attack. As a result, marginal portions of the till were left unsupported as the shale flowed out from below. Weakening and collapse followed, and thus individual bodies were launched on their downslope course.

In addition to the processes discussed above, it is possible that ice wedges, of the type described from Alaska by Leffingwell,[19] may have formed in frost cracks or in cracks due to collapse. Such ice wedges, if present, would have played some part in the initiation of landslide masses, either in process of detachment or during the early stages of movement, or both.

Movement of the landslide masses.—Initial movement of the landslide masses undoubtedly took place on slopes somewhat oversteepened by nivation and solifluction. On these slopes, direct gravitational settling on yielding, saturated clay was probably competent to effect movement. In addition, ice wedges between the till masses and the clay may have been a contributing factor, as noted above.

As the landslide masses moved down onto the gentler slopes below, the tangential component of gravity became weaker and the retarding effect of friction became greater. A point was quickly reached at which gravity was powerless to effect further movement unless some other process intervened to produce, temporarily, a steeper working gradient. The expansive force of frost constituted this additional process. Further movement was intermittent and consisted of alternating periods of frost heaving and gravitational settling.

The nature of this intermittent type of movement was studied experimentally by Davison in the eighties. Using an inclined box of soil, he observed measurable heaving and settling of a marker

[19] E. De K. Leffingwell, "Ground Ice Wedges," *ibid.*, Vol. XXIII (1915), pp. 635–54.

"float" in the course of a single freezing-thawing cycle of 24 hours. He described the process as follows:

> The only direction in which expansion can readily take place is outward and perpendicular to the surface. Every particle of soil in the frozen layer will therefore be displaced from its original position along the line of the normal (or perpendicular) to the surface of the soil; . . . . The more intense and lasting the frost, the thicker will be the frozen layer, and the greater the displacement of the surface particles.
>
> On the recurrence of warmer weather, the interstitial ice will be melted, and in melting will contract, and the particles will return, not, as they came, along the normal to the surface, but in a direction nearly vertically downwards owing to their weight; not quite vertically, however, because the adherence of each particle to its neighbors . . . . tends to bring it back to its old position. The particles will thus after every frost and thaw occupy a lower position down the slope than they did before, and the whole outer layer of the soilcap will in this way creep slightly downwards.[20]

Although the sliding on Canjilon Divide differed greatly in scale from the setup used by Davison, the process is believed to have been essentially similar. The periods of freezing and thawing, however, were probably seasonal rather than diurnal. The settling path of a large landslide mass probably deviated less from the vertical than that of the smaller particles studied by Davison, and may even have veered to a downslope direction. Settling may have been facilitated somewhat by repeated undermining due to nivation and solifluction.

Additional light on the nature of the heaving process is provided by Taber's recent experimental work.[21] Taber found that "freezing of soils often results in heaving much greater than that which can be attributed to change in volume of the water present." This occurs where the soils behave as open systems, allowing water to be drawn in by capillary action to build up definite layered segregations of ice. This effect is most marked in materials whose component particles are smallest, such as clay. The weathered shale of the Canjilon Divide area thus was most favorable for vigorous frost heaving, and the zone of contact between the till and the shale may have been especially favorable for the development of ice layers.

[20] C. Davison, "On the Creeping of the Soilcap through the Action of Frost," *Geol. Mag.*, Dec. III, Vol. VI (1889), pp. 255–61.

[21] Stephen Taber, "Frost Heaving," *Jour. Geol.*, Vol. XXXVII (1929), pp. 428–61; "The Mechanics of Frost Heaving," *ibid.*, Vol. XXXVIII (1930), pp. 303–17.

Evidence of the competency of frost heaving to have lifted large masses of superincumbent material to a significant height is found in Taber's experiments and in observations on engineering structures. Taber concluded that "the maximum pressure developed by growing ice crystals in these experiments must have been well over 14 tons per square foot."[22] Aaron[23] states that "it is not uncommon for short sections of pavement to heave 10 or 12 inches above the adjacent surface." Gilkey[24] describes "some 16-ton concrete piers which were heaved this past winter by as much as 3 in.," in a Midwestern city. Wyckoff[25] notes that, in northern New York, "a brick wall giving an earth pressure of 2,000 lb. per sq. ft. was raised $\frac{3}{4}$-in. and several piers supporting columns and roof trusses were raised from $\frac{1}{2}$ to $2\frac{3}{4}$ in." Other examples might be cited,[26] but these are sufficient to indicate the strength of frost action under climatic conditions which do not even rank as periglacial.

The process of movement postulated for the landslide masses of Canjilon Divide may be summarized in the form of an equation:

$$R = QNE \sin S ,$$

where R = distance traversed in unit time; Q = a qualifying factor for deviation of the path of the settling mass from the vertical = ab/ac in the diagram, Figure 10; E = distance through which a given particle is moved by frost expansion, measured in a directional normal to the slope; N = number of freezing-thawing cycles per unit time; and S = angle of slope.

The actual rate of movement is problematical. In the foregoing equation there are too many unknowns to permit specific evaluation. It is reasonably certain, however, that the rate of movement

[22] "Frost Heaving," *loc. cit.*, p. 452.

[23] Henry Aaron, "Frost Heave in Highways and Its Prevention," *Public Roads*, Vol. XV (1934), pp. 10–16.

[24] H. J. Gilkey, "Freezing Ground Acts Like Hydraulic Jack," *Eng. News-Record*, Vol. LXXIX (1917), pp. 360–61.

[25] L. B. Wyckoff, "Some Observations on Effect of Frost in Raising Weights," *ibid.*, Vol. LXXX (1918), pp. 627–28.

[26] L. De B. McCready, "Action of Frost in Heaving Concrete Piers," *ibid.*, Vol. XCI (1923), p. 360; C. D. Norton, "Water Expansion in Ground Cause of Heaving in Winter," *ibid.*, Vol. LXXX (1918), p. 1058.

varied directly with the sine of the angle of slope. Available data on frost heaving suggest, furthermore, that the movement was very slow, probably of the order of from a fraction of an inch to a few inches per year. Even this slow rate, when continued for the thousands of years during which periglacial conditions prevailed, was quite adequate to account for the present distribution of the landslide masses.

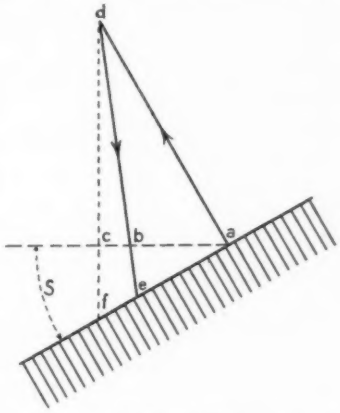

FIG. 10.—Diagram illustrating the movement of any given particle during the freezing-thawing cycle: *a* represents the initial position of the particle; *ad* (=*E*), the upward movement due to frost expansion; *de*, the return path followed by the settling particle as thawing takes place; and *e*, the final position of the particle; *df* represents the vertical path which, theoretically, would be followed by the settling particle if it were uninfluenced by factors other than gravity; *S* is the angle of slope of the surface.

At this point it may be asked why frost action, strong enough to effect notable displacement, did not disrupt the masses of till and destroy their identity. Undoubtedly there was some reduction of total bulk through surface disintegration. Taber's experiments show, however, that the presence of larger grains and particles tends definitely to inhibit the segregation of ice, and thus to produce greater resistance to frost attack.[27] Owing to this factor and to their higher position above the water table, the landslide masses probably functioned essentially as closed systems. Little additional water was drawn in, and the effects of such ice formation as took place were limited to interstitial adjustments which did not appreciably affect the structure as a whole.

Cessation of the movement.—With the gradual amelioration of climate attending deglaciation of areas farther north, frost action became feebler and finally was powerless to effect further movement of the till masses. The landslide bodies were left stranded wherever they happened to lie, to suffer only such further changes as might be

[27] "Frost Heaving," *loc. cit.*, pp. 435–41.

produced by normal weathering and erosion during the ensuing nonglacial interval.

CONCLUSIONS

The hummocky topography and the distribution of till on Canjilon Divide are satisfactorily explained only as being due to landsliding of integral bodies of till derived from a thick till mantle capping the remnants of a high-level erosion surface. It seems unlikely that this sliding could have taken place on the prevailingly gentle slopes unless aided by the intense frost action characteristic of periglacial conditions of climate, such as probably prevailed when the glaciers of the San Juan region lay at no great distance to the north. The size and the distribution of the various till masses with respect to one another and with respect to the different erosion surfaces can be explained best on the assumption of two widely separated periods of sliding, which are correlated tentatively with the Durango and the Wisconsin glacial stages of the San Juan Mountains in Colorado.

The initiation of the landslide masses is believed to have been effected mainly through undermining by nivation and solifluction. The movement of these masses on gentle slopes is interpreted as having been due to alternating frost heave and gravitational settling.

ACKNOWLEDGMENTS.—The writer wishes to express first his indebtedness to Professor Kirk Bryan, of Harvard University, whose lectures and discussions aroused interest in periglacial phenomena, and whose criticisms aided greatly in the preparation of this manuscript. Acknowledgment is due also to Mr. W. S. White and Mr. H. H. Mellus for able assistance in the field, to Mr. R. C. Goodwin for the use of photographs, and to Dr. C. F. Stewart Sharpe for a critical reading of the manuscript.

NOTES ON SHALLOW-WATER SAND STRUCTURES[1]

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ABSTRACT

Three widely contrasted types of sand structures are discussed in this paper—strand-line ripples, ridge and trough structures, and sand waves. The first is allied to but distinct from ordinary oscillation ripple marks. The third is analogous to ordinary, asymmetrical ripple marks but has crests rising above the troughs a foot or more instead of an inch or two. The ridge and trough structures are approximately symmetrical like oscillation ripple marks but are spaced at intervals of many feet instead of one or two or a very few inches. They are considered to be the product of the joint action of on-shore waves and undertow.

On a sandy beach a threadlike undulating ridge of sand[2] marks the highest reach of every wave in stormy weather. These delicate threads of sand, often not more than 3 millimeters in thickness, are sometimes seen in sandstone. They are known as wave marks and are familiar to every geologist. Another kind of strand-line structure which is characteristic of exceptionally quiet weather, just as wave marks are of a troubled sea, appears to be relatively unfamiliar to geologists if one can judge from the lack of references to it in the literature. These quiet-weather products of the strand line have been observed at Pelee Island in Lake Erie, on tidewater stretches of the St. Lawrence River below Montreal and elsewhere. Though occupying a zone of rather insignificant width, their occurrence just inside the water line would give them an interest if preserved in consolidated sediments, by supplementing wave marks in suggesting the relative positions of land and water at the time of their development. They appear to have been neglected in papers on ripple marks and other related sand structures.

Another type of sand structure of a very different order of magnitude which is extensively developed in shallow marine waters came to the writer's notice during the summer of 1933 when he partici-

[1] Published with the permission of the Director, Bureau of Economic Geology, Department of Mines, Ottawa.

[2] E. M. Kindle, "Recent and Fossil Ripple-Mark," *Can. Geol. Surv. Bull. 25* (1917), Pl. XIX*A*.

pated as a member of the International Geological Congress Excursion A5 in a cruise in Chesapeake Bay. The casual opportunities afforded by this cruise to observe various examples of these structures, which seem to have had less attention from geologists than they deserve, led him to revisit Chesapeake Bay, accompanied by Dr. C. H. Kindle, who assisted him in making a close examination of examples of these structures. The following notes relate to the aforementioned structures and to wavelike structures widely spaced which are characteristically developed in the Ottawa River.

STRAND-LINE RIPPLES

In quiet weather when the surface of Lake Erie is free from waves, shallow furrows may be seen along the west shore of Pelee Island in the coarse sand and fine gravel where the water is only a few inches deep. These structures run parallel to the shoreline and unlike ordinary wave-made ripple marks, which may extend a great distance from the shore, are limited to a very narrow belt seldom if ever more than a few feet wide. North of West Dock, Pelee Island, these structures, which may be called strand-line ripple marks, were seen as a pair of furrows in the sand 3–5 inches deep, spaced about 2 feet apart in water 4–10 inches deep. South of West Dock between the groins these strand-line furrows were observed in groups of four or more, irregularly spaced at from 8 to 20 or more inches. Their widely spaced crests distinguish them sharply from ordinary oscillation ripple marks.[3]

These structures are the product of the gentle backward and forward swing of the water at the shoreline, which is generally active in quiet weather or when a light breeze is moving off shore. The highly irregular spacing of strand-line ripples is probably correlated with the varying slope of the bottom near the shoreline. These structures, uncovered at low tide, were observed and photographed at Batiscan on the shore of the St. Lawrence 30 miles below Montreal (Fig. 1). These examples had been somewhat flattened and in places obliterated as the tide uncovered them. However, they clearly indicate the absence of ordinary oscillation ripple marks on both the land and the water side. Where the strand-line ripples have been observed,

[3] *Ibid.*, Pl. XIB.

the ordinary closely spaced oscillation ripples are commonly but not always absent. The wide spacing amounting often to several inches or more than one foot, the deep furrows, and the well-rounded symmetrical ridges when seen under water distinguish them from ordinary oscillation ripple marks.

FIG. 1.—Strand-line ripple marks in St. Lawrence River at Batiscan, Quebec

RIDGE AND TROUGH STRUCTURE

Another type of sand-bottom structure which is not built on the diminutive scale of strand-line ripples is developed very extensively around the shores of Chesapeake Bay (Fig. 2*a*). The unconsolidated and slightly consolidated sandy beds of Cretaceous and Tertiary age which form the shores of this bay furnish the waves an abundant supply of sand for new formations. The tidal currents and the waves together are redistributing the sands of the retreating bluffs which face the waters of Chesapeake Bay in many places. A discontinuous belt of clean sand skirts a considerable part of the Chesapeake Bay coast line as a result of active erosion of the shore and the contributions of numerous rivers.

North of the mouth of the Potomac River, Chesapeake Bay has a channel in which the greater depths run nearly 100 feet and in a few places reach 150 feet or a little more; but over the greater part of the area of the bay depths less than 50 feet prevail.[4] The five-foot contour often lies from a half-mile to a mile and a half from shore. Sections across the bay in the region of Chesapeake Beach would show the water increasing in depth from the eastern and western shores toward the middle of the bay at the rate of from 10 to 15 feet per mile for about 3 miles. At various points as far as 1,000 yards from the shoreline the maximum depth is 7 feet or less.

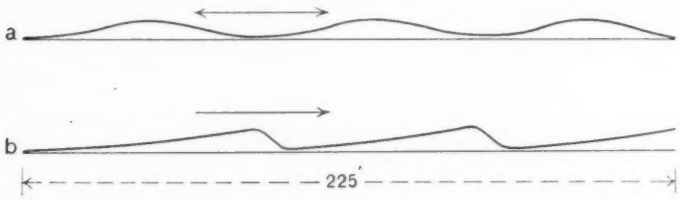

FIG. 2.—Diagrammatic illustrations of (*a*) ridge and trough structures and (*b*) sand waves. Single arrow indicates direction of river current and migration of sand waves. Double arrow pointing left shows direction of onshore waves; pointing right, direction of undertow currents—the two factors responsible for the nearly flat-topped ridges.

At certain points where observations were made the surface of the sand-zone bottom shows not a plane gently inclined toward the deeper off-shore water, such, for example, as one commonly finds along the southwest shore of James Bay, Ontario, where the bottom is generally clay or mud, but a billowy surface consisting of a series of widely spaced, nearly flat ridges and intervening troughs which trend parallel to the shoreline. These structures were seen well developed at Governor's Run, Calvert county, Maryland. Here the shallow troughs were found to be spaced at intervals of about 85 feet, the axes of the ridges rising 2 or 3 feet above the troughs. These symmetrical structures differ in shape as well as in origin from the meta-ripples[5] of rivers and estuaries which are asymmetric in form and are developed by strong current action at right angles to

[4] *U.S. Coast and Geodet. Surv. Charts 1225, 77,* and *78.*

[5] W. H. Twenhofel, *Treatise on Sedimentation* (Baltimore: Williams & Wilkins Co., 1932), p. 649.

the current direction. Neither can they be classed with the "fulls" of the English coast, which represent, according to Cornish, a single sand or shingle "ridge often with a lagoon behind it at high tide."[6] The writer's opportunities permitted him to observe these foreshore structures generally only in water less than 6 feet in depth, and the maximum width of the zone on which they occur is not known. Sometimes only three or four could be seen while ten were observed in one case, the axes of the troughs spaced at intervals approximating 60–100 feet.

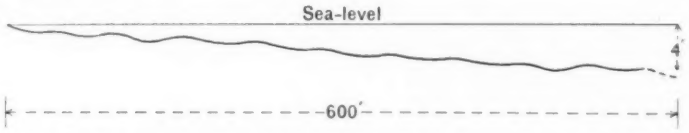

FIG. 3.—Sea-bottom profile showing ridge and trough structure at Fair Haven, Chesapeake Bay, Maryland.

At Fair Haven, Maryland, the bottom section consists of fine buff sand resting on dark-gray sand. A section about 600 feet in length at right angles to the shore was examined here. The first ridge structure was encountered about 300 feet from the high-tide shoreline. Ten of these structures, gently rising away from the shoreline for some 30 or 40 feet and as gently falling seaward for a like distance, were recognized in a distance of about 650 feet, the maximum depth reached beyond the outermost rise being $4\frac{1}{2}$ feet (Figs. 2*a* and 3).

At Governor's Run, Maryland, the ridges were found to be slightly steeper on the shore side than on the sea side.

Three aerial photographs published by Willis T. Lee[7] show submarine features of this type clearly and indicate that the parallel ridges occur in series which number in some cases as many as eight or ten ridges. These evenly spaced ridges are referred to by Lee in his explanations of the photographs as sand bars and are not distinguished by him in any way from the highly irregular underwater

[6] Vaughn Cornish, *Waves of the Sea and Other Waves* (Open Court Publishing Co., Chicago, 1910), p. 162.

[7] *The Face of the Earth as Seen from the Air* (New York: Amer. Geogr. Soc., 1922), Fig. 30, p. 46; Fig. 72; Fig. 76.

structures commonly called sand bars shown in some of his photographs from the air. The seaward limit shown by Lee's pictures affords evidence only of the penetrating powers of the lens used (Fig. 4); the actual width of the zone with subsurface ridges may be several times that disclosed by airplane pictures. One of Lee's photographs (Fig. 59) shows a combination of ridge and trough structure

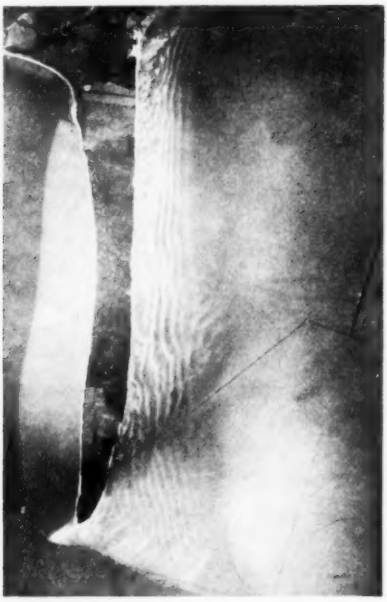

FIG. 4.—Photograph from the air at an elevation of about 10,000 feet showing ridge and trough structure east side of Stove Point Neck, south of mouth of Rappahannock River. Reproduced from W. T. Lee, *The Face of the Earth as Seen from the Air* (American Geographical Society of New York).

and sand waves; the latter, transverse to and partially obliterating the former, are doubtless the product of tidal currents.

Observations by James L. Kellogg on similar ridge and trough structures in the Gulf of Mexico on the Louisiana coast indicate a width from crest to crest in that area approximating the spacing observed by the writer in Chesapeake Bay. He states that "the

crests of these ridges were from ten to twenty feet across and they were twenty to thirty yards apart."[8] An interesting feature of Kellogg's description of these structures is the rapidity with which the character of the trough sediment changes. After wading "perhaps 1500 feet across slightly elevated ridges with hard bottom the same traverse was repeated two days later when the depressions were found to be three or four inches deep with very soft mud and the ridges free of mud. It thus appears that a complete change in the character of the material in the upper three or four inches of these trough structures may take place in 48 hours."[9]

Around the Bay of Fundy in Canada and the Basin of the Wash in England, the ebbing tide lays bare a belt of sea bottom in places more than 2 miles wide. Although sandy sediments abound in both of these basins which afford such exceptional opportunities for inspecting the sea bottom, the writer has seen nothing comparable to the ridge and trough structures of Chesapeake Bay. Their absence may be due to the strong tidal currents. The moderate range of the tides and consequent low velocity of the tidal currents in Chesapeake Bay are probably favorable elements in the perpetuation and development of these structures.

Structures of this type are found in the Great Lakes where the bottom is favorable. Three of the ridge and trough structures were found in front of the Fair Haven, New York, beach in Lake Ontario. West of the creek mouth at this beach the water deepens gradually to about $3\frac{1}{2}$ feet where a bar some 200 feet from shore rises about 1 foot above the trough, followed by a second trough and bar from which the water deepens quickly to more than 6 feet.

SAND WAVES

Structures analogous to current ripple marks, but of an entirely different order of magnitude, are developed in rivers which have an abundant supply of sand and a suitable gradient (Fig. 2*b*). River sand waves differ from the familiar sand waves of estuaries chiefly in being products of seasonal development. While the sand waves

[8] "Notes on Marine Mollusks of Louisiana," *Gulf Biol. Station Bull.* 3 (Cameron, La., 1905), p. 38.

[9] *Ibid.*

of tidal estuaries are modified or largely reconstructed with every change of the tide, the sand waves of an inland river are in an active stage of development only during its high-water stages.

In the Ottawa River, to which the present observations are limited, the annual spring rise usually reaches its peak during the last two weeks in May. Owing to the large lakes connected with this

FIG. 5.—Sand waves partly under water at Upper Duck Island, Ottawa River. Photograph from an elevation of 500 feet by Royal Canadian Air Force, Department of National Defence. Current direction indicated by arrow.

stream which serve as stabilizers, the spring rise-and-fall proceeds slowly. In most seasons active development of sand waves in the Ottawa River is limited to the months of May and June. Decline of the water-level and clearing of the water in July and August expose a wide belt of sand waves to direct observation on both sides of Upper Duck Island near the city of Ottawa. During the late summer the steep downstream fronts of the sand waves are stationary and remain unchanged below the water surface (Fig. 5). Above it the troughs tend to fill up and obscure the sand waves as a result of wind-

drifted sand. Observations extending over several years have shown that in years when the spring rise is exceptionally small and the current correspondingly weak during the high-water period, the sand waves fail to develop in the area shown by the photograph. This observation correlates well with the commonly accepted view that sand waves are the product of strong, overloaded currents.

There is considerable variation in the spacing of the crests of the Ottawa River sand waves developed near Ottawa. The spacing commonly ranges between 15 and 60 feet with downstream fronts at an angle near 33°. The long upstream side of the "wave" is nearly flat. Directly below the crest the depth of water abruptly increases by from 1 to $2\frac{1}{2}$ feet as compared with the depth over the sand-wave crest. Flying at 500–800 feet on a bright day in late summer when the water is clear, one can see with surprising distinctness the patches of straw-colored sand waves gleaming through the water, whereas from a canoe they could be seen only with difficulty or not at all. Observations from a hydroplane over the 35-mile stretch of the Ottawa River between Ottawa and Buckingham show the patches of sand on the river bottom to be nearly all marked by widely spaced sand waves. These tongues and patches of sand are not continuous for considerable distances but appear sharply outlined against the usual dark-gray clay or mud bottom of the river almost as clearly as small clouds stand out against a blue sky. The appearance of these structures as seen from the air is indicated in a photograph taken in late summer when the sand waves are partly uncovered (Fig. 5).

Abundance of sand is not the only requisite for the development of sand waves in a river. The writer has, during the low-water season, ascended the lower stretches of the Hamilton and Nascoupie rivers in Labrador without seeing these structures, although sand abounds in their channels as it does in no other rivers known to him. Instead of sand waves, the Nascoupie develops in places a peculiar type of pit and mound structure.[10]

[10] E. M. Kindle, "Unusual Type of Sandbar," *Pan-Amer. Geol.*, Vol. XXXIX (1923), pp. 15–16, Pl. 1.

REVIEWS

Historical Geology of the Antillean-Caribbean Region. By CHARLES SCHUCHERT. New York: John Wiley & Sons, 1935. Pp. 811; figs 107; paleogeographic maps 16. $10.

The Antillean-Caribbean region is one of the earth's "four-corners" where two continents and two oceans come close together. At times the continents have joined hands across it; at other times the oceans have become confluent over it. Morphologically it is composite, for it has acquired characteristics from each of these four major units. Naturally it is a region of deep basins and peculiar tectonic and volcanic arcs. Its history is of great significance both in its own development and in its relations to the development of North and South America. Biologically it is replete with problems.

The geologic literature covering the many different land areas which border the Gulf of Mexico and the Caribbean Sea is very scattered. To bring the essential features of the various fragments together and to fit them properly into their larger units and relations has been the aim of Professor Schuchert in this monumental work. Not only is the work a great compilation of facts and interpretations, made readily accessible to the reader, but the author has woven these together in a general synthetic study. Inevitably, however, because so many critically important portions of the region lie submerged beneath the several seas and not available for direct study, the hypothetical element looms somewhat larger in these expressed interpretations than in those of better-exposed regions. The author has naturally colored the picture in accordance with his views and preferred interpretations, but the guiding outlines are so clearly drawn that others may readily shade the colors to accord with their individual preferences. The critical reader will be grateful for the vast assemblage of significant facts and the carefully considered interpretations.

The volume is organized in four sections: I, "Introduction and Summary"; II, "The Three Middle American Basins"; III, "Biogeography"; and IV, "The Lands." Section I comprises an introductory chapter and three general summary chapters on the greater geologic features, diastrophism, and paleogeography. The greater geologic features are taken to be the Antillean geanticline, Caribbean mediterranean, Paria border-

land of South America, Antillean geosyncline, Gulf of Mexico "plate," Bermuda, Caribbee volcanic arc, Bahamas, and the Panama bridge.

Of prime importance and in some respects the central feature in the general interpretation is the Antillean geanticline. In southern Mexico the structural grain changes from northwest-southeast to approximately east-west, and thence the folded belt continues east or slightly north of east across Guatemala, Honduras, northern Nicaragua, and onward as a submerged, gradually tapering platform which reappears above the waves in Jamaica, Haiti, Puerto Rico, and the Virgin Islands. The folding, accompanied by granitic intrusions, occurred chiefly in late Paleozoic and late Cretaceous times with late Cenozoic epeirogeny and block faulting, called late Pliocene in Central America and late Miocene-Pliocene-Pleistocene in the Antilles. To the south is the Caribbean basin and to the north was the site of the old Antillean geosyncline, now in part the Antillean Sea and Cuba.

The Gulf of Mexico plate, an inheritance from Suess and strongly upheld by Schuchert, seems to the reviewer the least satisfactory element of the whole picture. The author's view is that this plate included the now very deeply sunken Gulf of Mexico along with Florida, the Bahama banks, and the northern portion of Yucatan, and that it constituted the "foreland" of the Greater Antilles and Central America.

> The subsiding of the Gulf of Mexico plate began as early as the Middle Silurian, but was of a gentle and intermittent nature throughout Paleozoic time. . . . . Then, following late Jurassic time, this long-persisting, slightly negative foreland began to change into a decidedly downwarping basin, and in the Cretaceous and Eocene it developed into a gulf that was far greater then the present one.

Today the Bahama banks, Florida, and northern Yucatan stand as broad platforms reaching beneath shallow seas far beyond the present shore lines, and the exceptionally thick Cretaceous and Tertiary sediments along the Louisiana and Texas Gulf coasts indicate that strong subsidence has occurred at least locally, but the evidence thus far presented does not seem to the reviewer necessarily to require the view that the middle portion of the Gulf of Mexico floor once stood flush with the Florida and Yucatan platforms in one vast plate from which it has since sunk to its present great depths. There may be other possibilities.

The independence of the Lesser Antilles, or Caribbee volcanic arc, from the Greater Antilles, or folded Antillean system, is emphasized. While the chain of Lesser Antilles curves gracefully from the eastern extremity of the folded Greater Antilles around to South America, the islands at least are

of volcanic origin. They rise as volcanic piles from a submarine swell which, however, is generally thought to be primarily a tectonic feature.

Of unusual interest in the vicinity of the Caribbees, though not geologically one of them, is the island of Barbados, treated under the heading of "Northern South America." Paria, the Paleozoic and Mesozoic northern borderland of South America, is believed to have curved to the east-northeast far enough to include Barbados as its most northeasterly outpost. The hydrographic charts indicate that it is not unreasonable to think that the land border once extended north of Colombia and Venezuela about as far as the present 1,000-fathom line, and its northern boundary could have cut across the site of the southern extremity of the subsequently formed Caribbee ridge to include Barbados. The east-northeast trend of the strong late-Eocene Barbados folds, transverse to the present north-south Barbados bank, seems to substantiate this, as do also the Barbados Eocene sediments derived from nearby land on the south. Later occurred what the reviewer would call a down swing of the outer margin of the continent rather than a foundering of Paria into the abyss, following which Barbados acquired the Oceanic series of interesting radiolarian and other unusual deposits. To subsequent upwarping may then be attributed the present slightly arcuate north-and-south Barbados bank, concave toward the west like the line of southern Caribbees, with which it is more or less concentric. The submarine Caribbee ridge which now laps upon the outer portion of the old borderland, together with the north and south Tobago trough, has since greatly modified the earlier relief features (see Fig. 106, p. 730).

Structurally Central America is in two main divisions: (1) The wide belt of older east-west trending folds between southern Mexico and northern Nicaragua, which Hill called the Antillean orogenic system; and (2) a chain of younger volcanoes running northwest and southeast, parallel to the Pacific shore line, through western Guatemala, Salvador, Nicaragua, and Costa Rica and into Panama. The two are quite distinct. From Lake Nicaragua south the volcanoes are perched on "the young isthmian link connecting Central and South America." This isthmian link is a tectonic arch which, at 6,000 feet beneath sea-level, has a base about 200 miles across in Costa Rica and Panama. Of its early history Schuchert says (pp. 27–28):

> The historical geology of the past and the biogeography of the present bring us no convincing evidence that North and South America were united by the Panama land bridge before Upper Cretaceous time. . . . . The best conjecture is that the Panama submarine ridge may have arisen during the late Jurassic diastro-

phism, and that in any event surmounting volcanoes had completed the land bridge before the close of the Cretaceous.

This seems entirely logical. In the mind of the reviewer, however, another question arises: By what route did the supposed migrations of shallow-water marine faunas between North and South America take place in early and middle Devonian times? Agreeing that the Caribbean was already a deep-sea basin and admitting the probability that the Caribbee submarine swell around the eastern margin of the Caribbean was of later development, are we to seek a shallow-water migration route via Panama as early as the Devonian, or is there some other solution of the problem? It is easier to ask the question than to answer it. The very important and interesting Cenozoic history of the Panama bridge and the marine portals which interrupted its continuity is more readily deciphered and is admirably interpreted from abundant available evidence presented in its appropriate place later in the book.

The early chapter on diastrophism states that the lands bordering the Caribbean mediterranean and the Gulf of Mexico, or lying within these areas, record at least five and possibly eleven times of diastrophic movement. Strongest and most widespread were orogenic revolutions in the late Paleozoic and late Cretaceous and the late Cenozoic epeirogeny (Antillean revolution) with much block faulting. These show more or less correspondence to the Appalachian and Laramide revolutions of the United States, and to the uplifting and block faulting of the Basin ranges and other parts of our western Cordillera which culminated in the late Pliocene.

Section II comprises relatively brief, condensed descriptions and discussions of the principal features of the three Middle American basins—the Caribbean and Antillean seas and the Gulf of Mexico. Elaborate detailed treatment is not attempted in this section because much further light is thrown upon these basins incidentally in the fuller treatment of the land areas which makes up most of the remainder of the book.

Section III, devoted to biogeography, contains first a valuable chapter on the principles of distribution of plants and animals, followed by another giving the detailed biogeographic evidence. In succession are considered plants, land snails, crayfishes, prawns, spiders, fishes, amphibians, reptiles, birds, and mammals. It is concluded that the Antillean faunas are too homogeneous throughout to have reached the islands by flotsam or jetsam dispersal. On the other hand, very few species are now common to Antillia and Central America, and this means long isolation. Nearly all

biogeographers are agreed that the Antillean life (birds excepted) is predominantly Central American in character, and that Central American life, in turn, is very largely of South American origin. The author finds unmistakable evidence of land migration from Central America to the Greater Antilles at two widely spaced times: first, during the Eocene and, second, in a much more limited way during the Miocene and early Pliocene. He recognizes, at these times, a land bridge from Honduras and Nicaragua across to Jamaica and Hispaniola, but he does not see evidence for a land bridge from Yucatan to Cuba and Hispaniola after Triassic time. Cuba, he thinks, was cut off from Haiti before the Honduras-Jamaica-Haiti bridge was broken down.

Section IV ("The Lands") is by far the greater part of the book, totaling 650 pages. By states, countries, and islands it treats successively the general geology from Florida around to Venezuela and includes also the Bahamas and Bermuda. Each of the thirty-six chapters is a mine of concisely presented information.

The book has very great value for two reasons: The mature geologist who will study it carefully and thoughtfully will obtain a clear picture and excellent understanding of the known geology and geologic problems of one of the most critical areas of the globe; he who needs trustworthy, up-to-date information upon any particular portion or problem of this extensive province can obtain it with the minimum of effort. Should he wish to delve further into the literature of any topic, the carefully selected and rather extensive Bibliography at the end of each natural group of chapters provides an excellent key to the original sources of information.

R. T. C.

Geology and Ground-Water Resources of the Island of Oahu, Hawaii. By HAROLD T. STEARNS and KENNETH N. VAKSVIK. Division of Hydrography, Territory of Hawaii, Bull. 1. Honolulu: Division of Hydrography, 1935. Pp. 479; figs. 34; pls. 33.

This report is an excellent treatment of the geology of the beautiful island on which Honolulu is located, and as such will be welcomed by all geologists who have visited Hawaii or are interested in it. Volcanic processes have played the chief rôle in the development of the island, but Oahu presents also some exceptionally interesting problems in geomorphology. What geologically instructed traveler has seen the precipitous Nuuanu Pali but who has wondered how it could have been formed! In this volume Stearns has outlined carefully the several successive stages in

what now seems a satisfactory interpretation involving amphitheater-headed valleys, undercutting of the Pali walls by emerging ground water, submergence by more than 1,200 feet which allowed the sea waves to batter back the divides between amphitheaters producing the straight cliff, deposition on the valley floors, and subsequent emergence. Likewise, the history of Pearl Harbor proves to have been much more complicated than earlier writers had supposed.

Because of the high permeability of the lava rocks and the flashy character of the streams the water supply of the island is obtained mainly from wells, tunnels, and springs. Some of the artesian conditions and problems are of much more than local interest. More than half of the report is devoted to a thorough consideration of the ground-water resources.

R. T. C.

Contribuicão para a geologica do estado de Goyaz. By GUILHERME BASTOS MILWARD. São Paulo: Escolas Profissionaes Salesianas, 1935. Pp. 98; figs. 3; map 1.

This work, written by the late Guilherme Milward in 1926 after two years work, is now published as the first of a set of commemorative volumes. It gives the results of a pioneer geological and physiographical survey, covering about 15,000 square miles of the state of Goyaz (Brazil). Two principal Mesozoic formations are distinguished, which are probably Cretaceous—the Bauru and the Batucatu—both principally sandstones with intrusives, which overlie various sedimentary and metamorphic Paleozoics and earlier gneisses.

FRANKLIN MACKNIGHT

Lake Huron Sheet, Map 155A: Publication No. 1553. 3d ed. Geological Survey of Canada, Ottawa, 1933. Scale: 1 inch to 8 miles (1/506,880).

The third edition of this well-known map shows a few corrections and additions over the earlier editions. Most notable is the addition of a marginal text on the geology and economic resources.

No fundamental change is made either in classification or in mapping. The sheet includes the classic Huronian areas along the north shore of Lake Huron and Cobalt, the Sudbury structure, the type Timiskaming area, and the type Killarney region. In short, the best-known and most complete pre-Cambrian section of the Canadian shield is here portrayed.

Lake Nipigon Sheet, Map 308A: Publication No. 2364. Geological Survey of Canada, Ottawa, 1935. Scale: 1 inch to 8 miles (1/506,880). Lat. 48°–52° N. and Long. 85°–90° W.

This map is a compilation of all available information on the geology of the area north of Lake Superior and around Lake Nipigon.

The bedrock includes an early pre-Cambrian complex consisting of mica-schist and less-altered sediments variously designated Coutchiching, Windigokan, and "mica schist," of lavas and pyroclastics, in part marked Keewatin and in part undesignated, and of large granite intrusives. The late pre-Cambrian consists of low-dipping to flat sediments (Sibley) and lavas (Osler), together with the extraordinary Nipigon diabase sheet and small quartz-porphyry intrusives. Paleozoic (Silurian) of the Hudson Bay lowland is present in the northeast corner of the map area.

Rouyn-Bell River Area, Map No. 328A: Publication No. 2404. Geological Survey of Canada, Ottawa, 1936. Abitibi and Temiscamingue counties, Quebec. Scale: 1 inch to 4 miles (1/253,440).

This map is a compilation of surveys by the Geological Survey of Canada and the Bureau of Mines, Quebec. With the exception of a small area of Cobalt (Huronian), all the rocks are early pre-Cambrian. These are Keewatin lavas and tuffs with some possible post-Keewatin sediments, Timiskaming conglomerates and associated sediments and paragneisses, and a little post-Timiskaming basic lava. A variety of intrusives, mainly granite, are mapped. Of special interest are the gabbro dikes, in part late pre-Cambrian in age, one of which is essentially continuous for over 80 miles.

Mine Center Area: Rainy River District, Map No. 334A: Geological Survey of Canada, Ottawa, 1936. Scale: 1 inch to ½ mile (1/31,680).

This map records the result of recent detailed field work by T. L. Tanton in the classic Seine River and Shoal Lake area just east of Rainy Lake. The rocks involved are Coutchiching, Keewatin, and Seine, with a variety of post-Seine intrusives including some anorthosite and much granite.

F. J. PETTIJOHN